Forest Soil Temperature Manual

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February 1990

ECONOMIC & REGIONAL DEVELOPMENT AGREEMENT

Canada

BC

91-02347/

Funding for this publication was provided by the Canada-British Columbia Forest Resource Development Agreement - a five year (1985-90) $300 million program cost-shared equally by the federal and provincial governments.

Canadian Cataloguing in Publication Data

Stathers, Robert John, 1957-
Forest soil temperature manual

(FRDA report, ISSN 0835-0752 ; 130)

Issued under Canada-BC Forest Resource Development Agreement.
Co-published by B.C. Ministry of Forests.
"Canada/BC Econonic & Regional Development Agreement."
Includes bibliographical references.
ISBN 0-7726-1183-1

1. Soil temperature. 2. Forest soils. 3. Conifers - Soils. I. Spittlehouse, David Leslie, 1948- II. Canada. Forestry Canada. III. British Columbia. Ministry of Forests. IV. Canada-British Columbia Forest Resource Development Agreement. V. Canada/BC Economic & Regional Development Agreement. VI. Titie. VII. Series.

SD390.S72 1990 631.4'36 C90-092216-8

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Province of British Columbia

This is a joint publication of Forestry Canada and the British Columbia Ministry of Forests.

Produced and distributed by the Ministry of Forests, Research Branch.

For additional copies and/or further information about the Canada-British Columbia Forest Resource Development Agreement, contact:

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506 West Burnside Road
Victoria, B.C. V8Z 1M5
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ACKNOWLEDGEMENTS

The authors gratefully acknowledge the reviews of Dr. M.D. Novak and Dr. T.A. Black, University of British Columbia, and L.J. Herring, W.R. Mitchell, O. Steen, P.M. Osberg, C.F. Thompson, A. Vyse, and P. LePage of the B.C. Ministry of Forests. We also thank T.A. Black, R. Korol, T. Letchford, D.A. Lloyd, and P.M. Osberg for the use of their unpublished data.

PREFACE

Environmental conditions are important in determining the survival, growth and distribution of tree species. However, information about the forest soil thermal regimes in British Columbia is limited. This has prompted forest ecologists and silviculturalists to investigate the atmospheric and soil climate of forests and clearcuts throughout the province.

The purpose of this manual is to provide foresters and other resource managers with a reference to aid them in the collection and interpretation of soil temperature information. To this end, the manual describes how silvicultural treatments can modify soil thermal regimes, and places emphasis on improving the soil thermal regime for conifer seedlings growing in forest clearcuts in British Columbia. It also describes the variations in temperature with time and with depth in the soil profile, the physical factors which influence soil temperature, the methods and instruments used to measure soil temperature, the procedures for interpreting soil temperature measurements, and the methods of classifying soil thermal regimes.

The manual consolidates material from a large number of texts and papers. A wide selection of these are listed in the bibliography for readers who are interested in greater detail.

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1 INTRODUCTION

The soil thermal regime at a site can have a strong influence on the survival and growth of conifer seedlings. Soil temperature directly affects the seedlings' physiological processes such as root growth (Figure 1.1) and water uptake (Figure 1.2). It indirectly affects seedlings through many of the physical, chemical and biological processes in the soil profile, for example, the weathering of soil minerals, nitrogen mineralization, organic matter decomposition, and microbial activity.

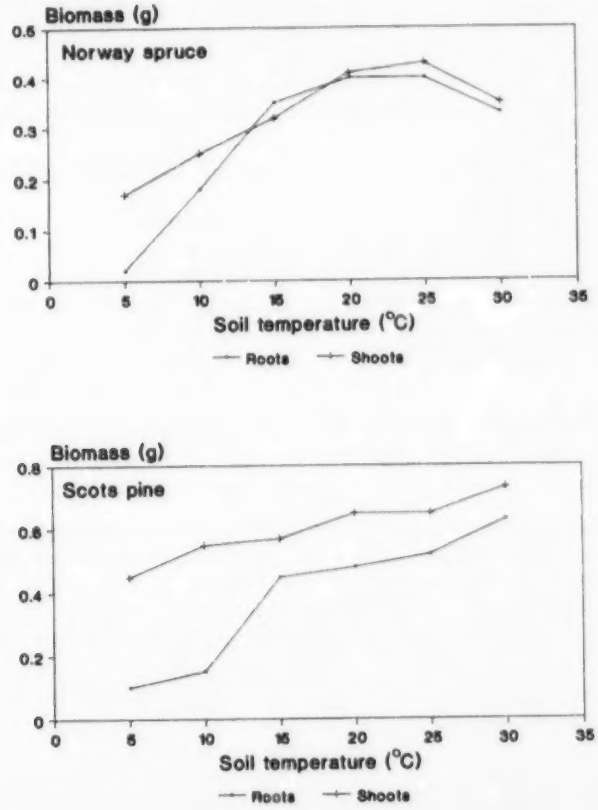

FIGURE 1.1. Effect of soil temperature on root and shoot growth of Norway spruce and Scots pine seedlings. (Adapted from Örlander *et al.* 1990.)

Chemical and biological processes usually have an optimal temperature range where the reaction rate is maximal, and upper and lower temperatures beyond which reactions do not occur. Many biological processes have an optimum operating temperature between 20 and 30°C, and are strongly limited by temperature below 10°C and above 40°C. Soil temperatures at forest sites are frequently well below optimum values. As a result, site preparation treatments which improve the soil thermal regime can significantly influence seedling growth and survival by altering biological reactions.

Soil temperature is determined by site location, atmospheric weather conditions, ground cover, and the physical properties of the soil profile. The location of the site (latitude, elevation, slope, and aspect) influences the receipt of solar energy. Atmospheric weather conditions determine the supply of radiation, heat and water to the ground surface. Vegetation and snow cover modify the effects of atmospheric

weather by shading and insulating the soil surface. In turn, the thermal and hydraulic properties of the soil profile govern the soil's response to the changing supply of heat and water.

Low and high soil temperatures are known to cause reforestation problems. Understanding the processes that affect soil warming and heat transfer should help foresters determine the thermal limitations at a site and select silvicultural treatments to overcome these limitations. This publication is intended to aid this process.

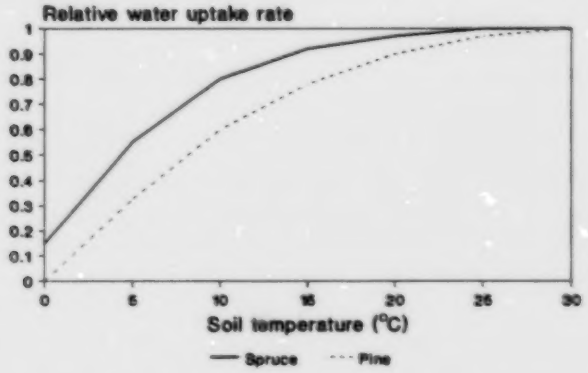

FIGURE 1.2. Effect of soil temperature on water uptake of Norway spruce and Scots pine seedlings. (Adapted from Örlander *et al.* 1990.)

Section 2 summarizes how site preparation can influence the soil thermal regime, and uses examples from studies in British Columbia and elsewhere. Section 3 presents the physical principles involved in heating and cooling the soil, and describes how soil temperature varies with depth on an annual and diurnal basis. This detailed analysis should help foresters in applying the information presented in Section 2 to other sites. Subsequent sections cover the measurement and interpretation of soil temperature data; and these are followed by a bibliography and a glossary of terms. The appendices present methods of calibrating and installing soil temperature sensors; a detailed description of the principles of heat flow in the soil; and the calculation of soil thermal properties.

2 MODIFYING THE SOIL THERMAL REGIME BY SITE PREPARATION

Harvesting methods and site preparation treatments can be used to improve the adverse environmental conditions that occur in the seedling root zone. These may involve lengthening the growing season by increasing the number of days when root zone temperatures exceed the minimum required for growth, or ensuring that lethally high root zone soil temperatures do not occur. These objectives can be accomplished to varying degrees through the use of management practices which alter energy receipt at the soil surface or modify the physical properties of the soil profile (Sutton 1989; Örlander *et al.* 1990; Spittlehouse and Stathers 1990).

Usually treatments that modify the soil thermal regime also influence other aspects of the seedlings microclimate. It is, therefore, necessary to determine the overall effect that a treatment may have. Although there may be additional beneficial effects, such as an increased amount of light available for seedling photosynthesis, there may also be an increased risk of creating detrimental conditions, such as exposure to high surface temperature, increased soil evaporation, soil compaction, and removal of soil nutrients. As well, such factors as cost and equipment availability must be considered.

Site preparation treatments that improve soil warming are usually most beneficial in the wetter, cooler subzones in British Columbia. At drier, warmer sites, greater gain can be made by improving the aerial environment of the seedling. At hot sites, treatments that reduce high surface temperatures can be beneficial. Most of the examples presented below are for the wetter cooler sites. The biogeoclimatic classification (Pojar *et al.* 1987; Meidinger and Pojar 1990) and physical aspects of the site (Canada Dept. Agric. 1972; Canadian Soil Survey Comm. 1978; Luttmerding *et al.* 1990) are given where available.

2.1 Altering the Energy Received at the Soil Surface

Solar and longwave radiation and the heat in the air provide the energy to warm the soil. However, the amount of warming or cooling depends on the vegetation cover, the partitioning of the energy received at the surface, and the transfer and storage of heat in the soil profile.

2.1.1 Changing the solar radiation balance

Solar radiation has a major influence on soil temperature. Consequently, management practices that alter solar energy receipt at the ground surface are effective methods of modifying the soil thermal regime. **Removing competing vegetation**, or **thinning a stand** will increase the solar irradiance at the ground surface and consequently increase the rate of soil warming. Conversely, **shading** can reduce high soil surface temperatures, reduce the soil evaporation rate, and cool the soil profile. Usually root zone soil temperatures are not sub-optimal at sites where shading treatments are applicable.

Allowing more radiation to reach the soil surface may not provide sufficient warming of the soil. There may be a need to alter the solar radiation absorption properties of the surface. This can be achieved by **altering the albedo of the surface** (the ratio of reflected to incident solar radiation).

Soils in cool, moist environments often have a thick surface organic layer. This layer must be removed to provide adequate soil warming. Altering how heat enters and warms the soil is discussed in Section 2.2.

Herbicide

The increase in soil temperature obtained using a herbicide treatment depends on the depth of the organic layer. Figure 2.1 shows the effect for a relatively thin organic layer. Warming was improved through the whole of the growing season. The herbicide treatment averaged 3-5°C warmer than the untreated area. Removal of the organic layer by scalping further increased soil temperature. The influence of thicker organic layers on warming is discussed in Sections 2.2.2 and 2.2.3.

Removing the vegetation has its greatest effect on the aerial environment of the seedling. There is an increase in the light available for seedling photosynthesis, and a reduction in vegetative press damage under the snow pack during the winter (Spittlehouse and Stathers 1990).

Shade cards

Natural or artificial shade sources are usually used in hot, dry environments. They shade the soil surface around the seedling root collar and some of the foliage. The use of natural shade sources involves planting behind large stumps or rocks, or beneath fallen logs (Hungerford and Babbitt 1987). Artificial shading includes the use of shade cards of various colours and materials and of styrofoam cups. Shade cards only affect soil temperatures near the surface behind the shade card. The effect of the shade card on temperatures deeper in the seedling root zone is

negligible because heat transfer from the surrounding warmer open areas overwhelms the effect of the shade card. However, reduction of high near-surface temperatures is usually all that is necessary.

The effect of a shade card on surface temperature on a steep, south-facing clearcut during a hot summer day is shown in Figure 2.2. The shade card was approximately 20 by 30 cm and oriented to the southwest. Surface temperatures near the root collar exceeded 68°C on the untreated open surface during the early afternoon, while at the same time they were up to 30°C lower behind the shade card. In the late afternoon the surface temperature behind the shade card increased for a few hours as the sun shone on the ground behind the shade card, though the temperature did not reach that of unshaded ground. Data from Oregon (Figure 2.3) show that these differences can be maintained through the summer.

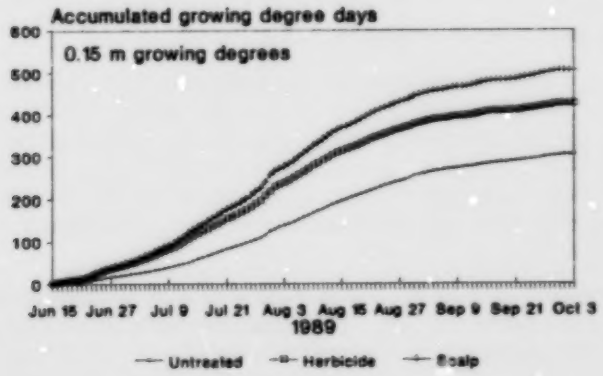

FIGURE 2.1. The effect of herbicide and scalping treatments on soil temperature at the 15 cm depth in the Engelmann Spruce - Subalpine Fir zone (ESSFwc2) near Clearwater. Response is expressed in growing degree days calculated with a 5°C base temperature. (D.A. Lloyd, 1989, B.C. Forest Service, unpublished data.)

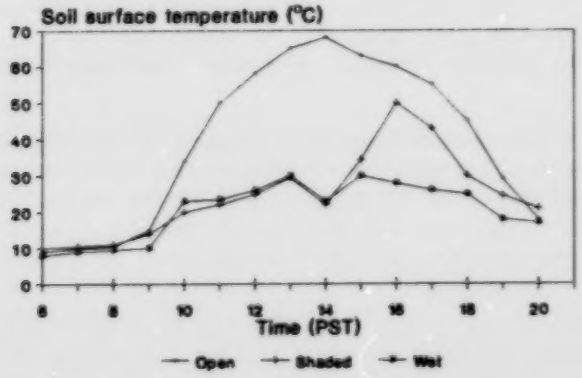

FIGURE 2.2. Surface temperatures on a clear hot summer day adjacent to Douglas-fir seedlings located in unshaded, dry soil, in unshaded, irrigated soil, and behind a shade card. The site is on a south-facing clearcut, at 1150 m, on gravelly soil in the Coastal Western Hemlock zone (CWHmm2), east of Port Alberni, B.C. (Adapted from Stathers 1983.)

Temperature was greatly reduced in the open when the surface was wet (Figure 2.2). This was the result of much of the absorbed energy being used to evaporate water rather than to warm the soil.

Seidel (1985) has shown that seedling heat stress is determined by the duration and intensity of high temperature. The shade cards (and shelterwood) shown in Figure 2.3 significantly reduced the duration of high temperatures (Childs and Flint 1987).

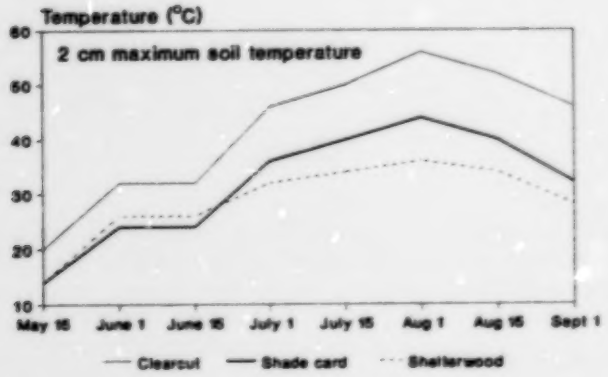

FIGURE 2.3. Daily maximum soil temperature at 2 cm in a clearcut, behind a shade card, and beneath a shelterwood in a Douglas-fir stand, in southwestern Oregon, on a stoney, sand soil. (Adapted from Childs and Flint 1987.)

In the studies shown in Figures 2.2 and 2.3, the shaded seedlings had a higher survival rate than non-shaded seedlings. The reduction in seedling mortality was not necessarily only a result of reduced surface temperatures. Shading the surface also decreased the soil evaporation rate, which conserved moisture and reduced drought stress (Childs and Flint 1987; Livingston and Black 1987).

Different types of shade cards are available, but their effectiveness has been variable (FIR 1986). They appear to be most suitable for steep, south-facing slopes in dry areas where frequent clear-sky conditions during the summer result in a high risk for high surface temperatures. Higher surface temperatures are also more likely to occur where organic or stony surface soils occur in these high risk areas.

Partial cutting

A partial cut of the forest canopy provides some shade and lowers soil temperature (Figures 2.3 and 2.4). Figure 2.4 shows mid-afternoon soil temperatures on a clear summer day for adjacent stands of interior Douglas-fir near Kamloops, B.C., with canopy crown covers of 100, 50, and 0%. In the clearcut, near-surface soil temperatures were high enough at times to be potentially damaging to seedlings. Where the dense stand canopy crown cover had been reduced by 50%, near-surface temperatures were more favourable for seedling regeneration. A similar situation occurred under the shelterwood shown in Figure 2.3.

Other benefits of partial cutting included a reduction in evaporative demand on the seedlings, and an increase in frost protection (see Section 2.1.2).

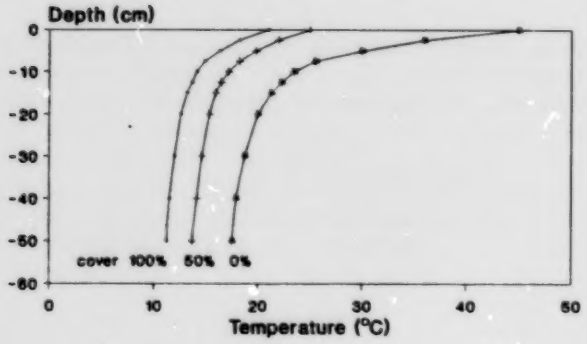

FIGURE 2.4. The effect of partial cutting on mid-afternoon soil profile temperatures in interior Douglas-fir stands with canopy crown covers of 100, 50, and 0%. (R. Korol 1983, Univ. British Columbia unpublished data.)

Altering the surface albedo

Modifying the albedo (by changing the colour of the surface) alters the amount of solar radiation absorbed by the surface. Blackening a surface by burning produces a low albedo and can increase surface temperatures by more than 10°C (Figure 2.5). Burning a surface also affects temperature through changes in soil thermal characteristics (see Section 2.2.2). In nursery or agricultural applications, the surface is lightened (e.g., with lime) to increase the albedo and reduce the soil temperature. It is darkened with charcoal to decrease the albedo and increase the soil temperature.

2.1.2 Changing the longwave radiation balance

The longwave (or thermal) radiation regime is modified by changing the structure of the vegetation canopy. A canopy reduces longwave radiation emission from the ground surface toward the colder night sky because much of the longwave radiation is absorbed by the vegetation canopy and subsequently re-emitted back toward the ground surface (Stathers 1989; Spittlehouse and Stathers 1990). However, this same canopy intercepts solar radiation during the day and the net effect is to lower soil temperatures (Figures 2.3 and 2.4). The best compromise is probably a moderately sparse canopy (about 50% cover) which lets some light penetrate during the day, but also reduces longwave radiative emission from the ground at night.

The reduction in night-time longwave loss decreases the risk of radiation frosts during the growing season. Studies in the United States and British Columbia obtained up to a 4-fold increase in the number of frost-free days at the ground surface in a selection cut than in a clearcut (Hungerford and Babbitt 1987; Stathers 1989).

2.2 Altering the Soil Heat Flux and Soil Warming

Allowing more radiation to be absorbed by the soil surface may not provide sufficient warming of the soil. There may be a need to change the thermal characteristics of the surface layer through mulching the surface, exposing mineral soil, tilling the soil, or creating mounds or trenches. These changes alter the depth of penetration of heat and the temperature increase caused by this heat. (See Appendix A for the relationship between heat and temperature.)

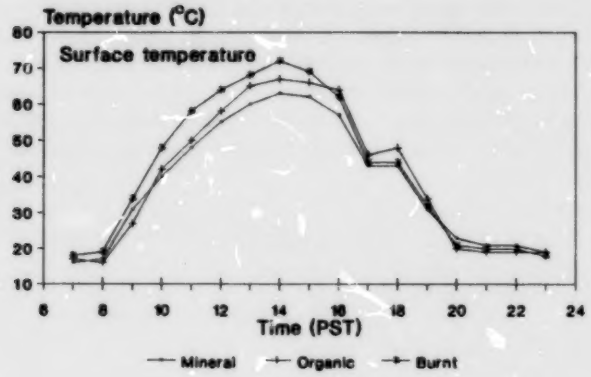

FIGURE 2.5. Soil surface temperatures of burnt, mineral, and organic surfaces on a southwest aspect, 55% slope, 700 m, gravelly, sandy loam soil, Coastal Western Hemlock zone (CWHvm2), UBC Research Forest, Haney, B.C. The 0.3-m temperature averaged 18°C for all treatments. (Adapted from Ballard *et al.* 1977.)

2.2.1 Mulching the ground surface

An organic layer acts as an insulating layer (mulch) which reduces heat transfer and storage in the profile. Organic material has a much lower thermal conductivity than mineral soil, particularly when dry, and is less effective in transferring heat into the profile. Most of the solar radiation absorbed by an organic surface is dissipated as sensible heat to the atmosphere. Thus, a thick organic layer results in low underlying soil temperatures (Figure 2.6) which change slowly over the growing season. Wet organic material also acts to suppress mineral soil warming because it has a high volumetric heat capacity.

The low heat capacity of an organic layer causes high surface temperatures on exposed surfaces on sunny days. These can potentially damage the root collar of seedlings. In addition, the lower amount of heat stored in organic surfaces means that at night they can radiatively cool to lower temperatures than mineral soil, thus increasing the risk of frost (Stathers 1989).

When dry, a mulch conserves moisture by reducing evaporation from the underlying mineral soil (Black *et al.* 1987; Childs and Flint 1987; Flint and Childs 1987). Benefits from this mulching are most likely to be found in drier environments where low root zone soil temperatures are probably not the most limiting factor to the growth and survival of seedlings.

Herbicide

A herbicide treatment creates a mulch of dead foliage. This dead foliage and the reduced amount of live vegetation cover after treatment results in more solar radiation reaching the ground surface. The net effect is usually to increase soil temperature (Figures 2.1, 2.7, and 2.8). However, this increase in temperature is often less than half that obtained by removing the surface organic layer. Herbicide treatments often have a greater effect on seedling survival and growth by increasing light for photosynthesis, and reducing moisture competition and vegetation press.

Plastic mulch

Plastic mulches, often used in nurseries and agricultural operations, have the opposite effect to that of vegetative or soil mulches. A clear plastic mulch transmits solar radiation. Since less evaporation and turbulence occur beneath a plastic mulch, heat accumulates and raises the temperature at the ground surface, causing an increased flow of heat into the soil (Hanks *et al.* 1961). Clear plastic mulches are more effective than black plastic mulches because they transmit solar radiation rather than reflecting it. So far, mulches have seen limited use in forestry because of their relatively high cost.

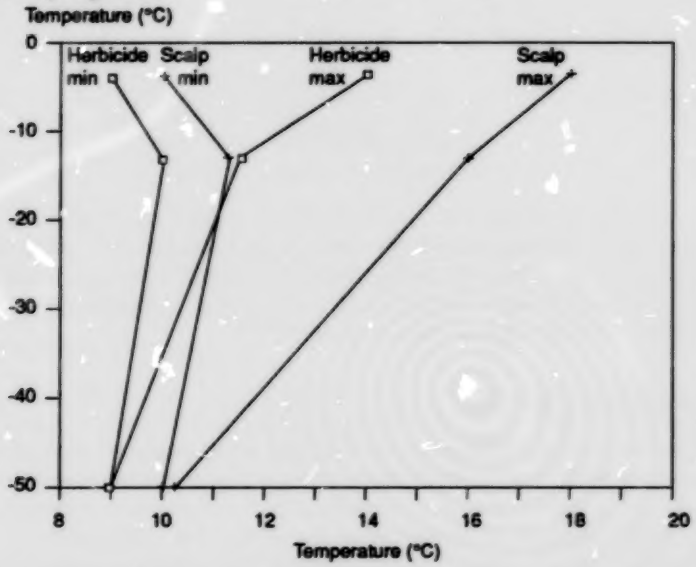

FIGURE 2.6. Soil temperature under an organic mulch and bare soil surface in the Engelmann Spruce-Subalpine Fir zone (ESSFxc) north of Kamloops, B.C. The mulch was produced by a herbicide treatment that left a 5 cm thick organic layer over the mineral soil. Blade scalping was used to expose the mineral soil. (Adapted from Black *et al.* 1987.)

2.2.2 Exposing the mineral soil

Removing the surface layer of organic matter or incorporating it into underlying mineral soil increases soil temperatures in the seedling root zone (Figures 2.6 to 2.9). This is partly because of the improved soil heat storage and heat conduction properties of the surface mineral soil layer. Mixing organic matter into mineral soil also increases the water holding capacity in the seedling root zone and reduces the likelihood of seedling water stress during periods of drought.

Exposing mineral soil can have a detrimental effect at drier sites by increasing the rate of soil surface evaporation. Loss of the soil nutrient reserve in the organic layer and soil compaction during the treatment of the site can also have detrimental effects. Patch scalping may resolve these problems to a certain extent. The size of patch required depends on a number of factors. Larger patches are necessary where competing vegetation could shade or rapidly re-invade the treated area.

Prescribed fire

Burning removes the vegetation which shades the surface and reduces the depth of the organic layer. It also decreases the albedo of the ground surface, thus increasing the amount of solar radiation absorbed by the surface. Removing or reducing the depth of the organic surface layer reduces the mulching effect of this layer. This can result in a 10°C increase in the surface

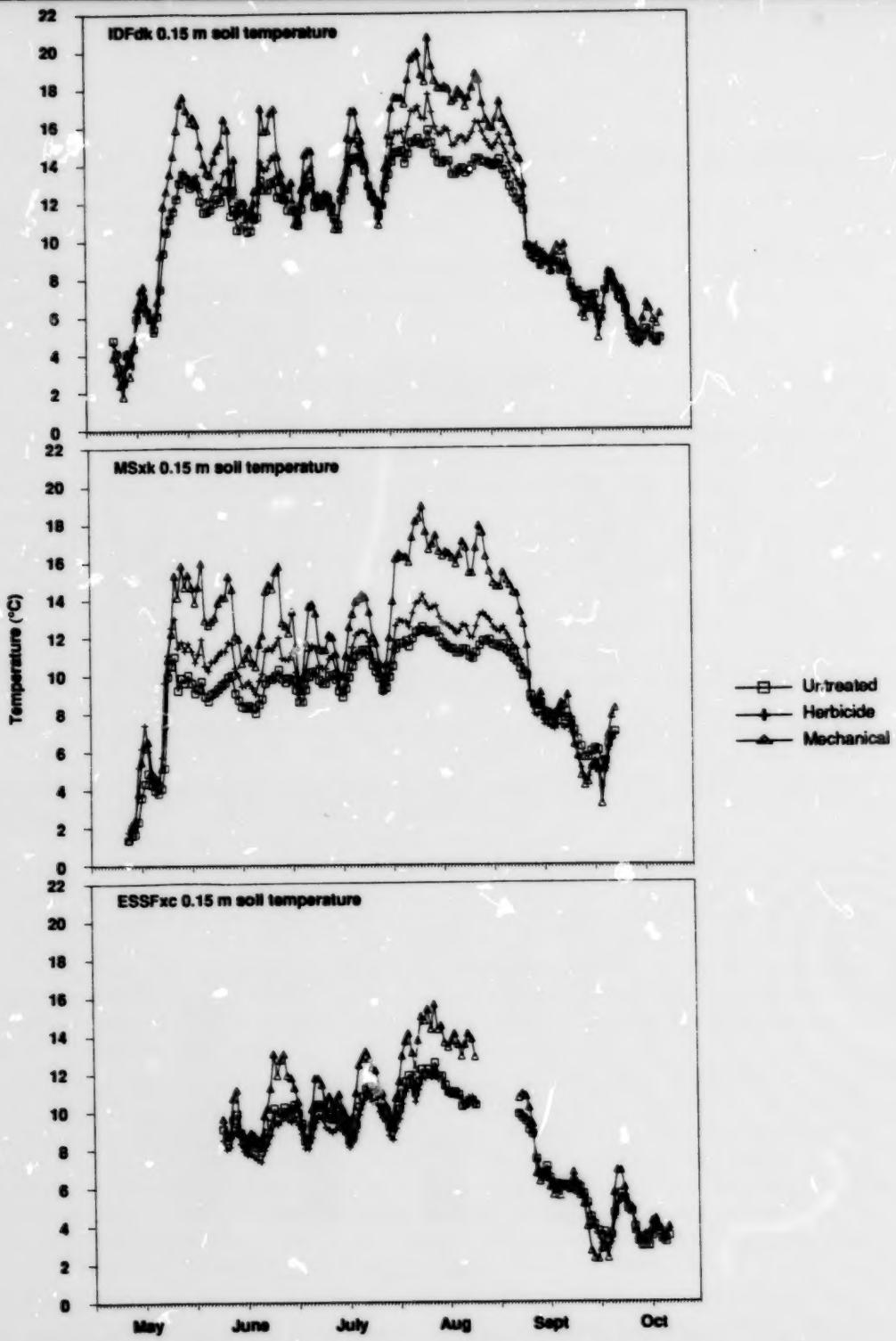

FIGURE 2.7. Mean daily 15 cm soil temperature for untreated, and herbicide and mechanical site preparation treatments in three subzones near Kamloops, B.C. The upper graph is for the Interior Douglas-fir zone (IDFdk1), the middle is for the Montane Spruce zone (MSxk), and the lower is for the Engelmann Spruce-Subalpine Fir zone (ESSFxc). All sites had pine grass and herbs as the dominant competing vegetation. The mechanical treatments (scalping and ripping) had the same temperature. (Adapted from Black *et al.* 1987.)

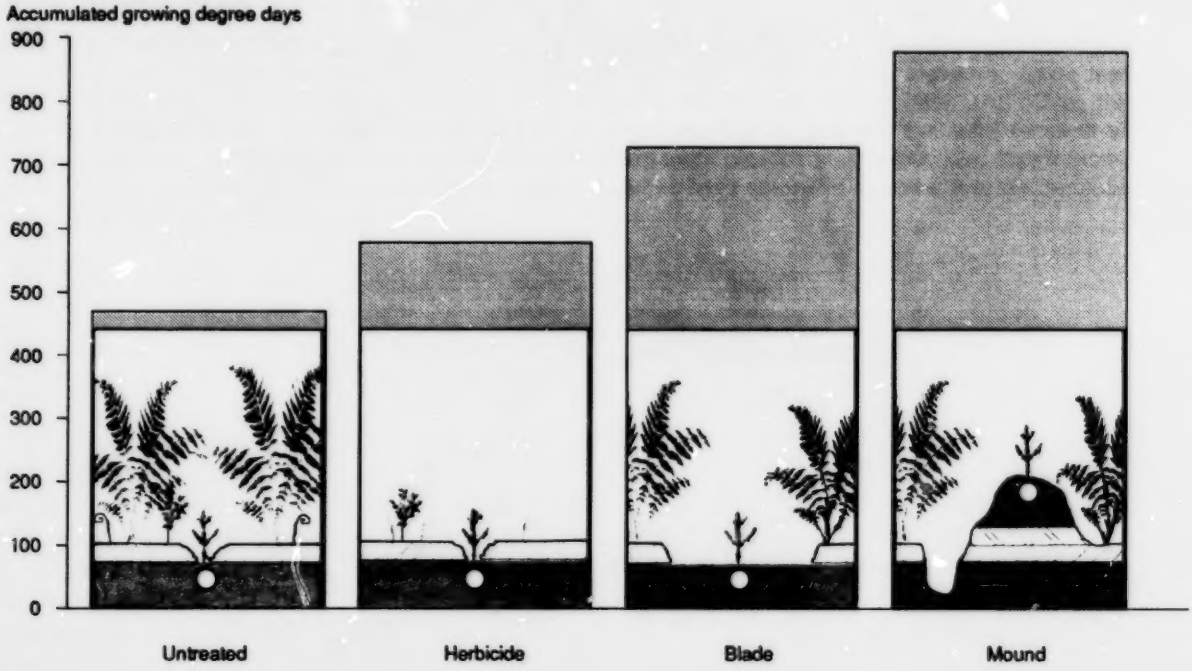

FIGURE 2.8. The effect of site preparation treatment on the accumulated growing degree days at the 10 cm depth in the Sub-Boreal Spruce zone (SBSvk), 1100 m, southeast of Prince George, B.C. (Adapted from Draper *et al.* 1988.) Ladyfern is the dominant competing vegetation. The soil is a silty loam.

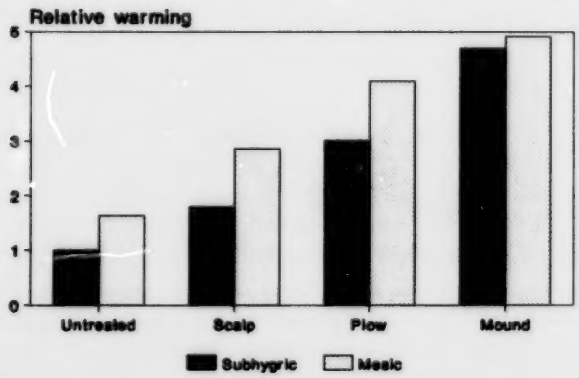

FIGURE 2.9. The effect of site preparation treatments on soil warming in subhygric and mesic soil moisture regimes in the Sub-Boreal Spruce zone (SBSmc2), 1010 m, north of Houston, B.C. The comparison uses a relative scale of accumulated heat units above 10°C in July, with the untreated subhygric regime measurements as the reference value. (Adapted from Macadam 1989b.)

temperature (Figure 2.5), and a 5°C increase at 10 cm during the daytime on sunny days (Hungerford and Babbitt 1987). However, if most of the organic material is not removed, seedling root zone temperatures can remain sub-optimal because the remaining organic material can act as an effective insulating layer.

The main benefits of burning are the improvement of the light regime, decreased water use by competing vegetation, and improved nutrient availability (Macadam 1989a).

Patch scalping

Mineral soil exposure involves the removal of both shading vegetation and the insulating surface organic layer. The importance of the latter can be seen in Figures 2.6-2.9. Figure 2.7 illustrates this for the same treatments over a range of subzones near Kamloops, B.C. (Black *et al.* 1987). The scalping and ripping treatments (grouped as mechanical) produced similar temperatures, averaging 2-6°C, greater than the herbicide treatment and untreated area. There is a general decrease in temperature, independent of treatment, with increasing elevation. This is due to a decrease in the snow-free period (less time for heating the soil) and an increase in soil moisture (increased soil heat capacity) with elevation.

Figure 2.1 shows similar results to those in Figure 2.7, in a cold wet environment (D.Lloyd, 1989, B.C. Forest Service, unpublished data). The site has a covering of herbs and a 5 cm thick surface organic layer. Accumulated growing degree days, or heat units (see glossary for definition), are greatest where mineral soil was exposed (blade scalped). About half of the temperature increase over the untreated area was obtained by removing the organic surface. The other half was obtained by removing the shading vegetation and allowing more solar radiation to reach the soil surface. Size of scalp is important. A patch scalp at this site had temperatures similar to those in the herbicide treatment. Although the mineral soil was exposed, the surrounding vegetation shaded much of the surface, reducing the amount of energy available to warm the soil.

The effect of a range of treatments on the soil temperature regime in a cold, wet environment near Prince George, B.C. (Draper *et al.* 1985, 1988) is shown in Figure 2.8. The site has a dense canopy of ladyfern and a 0.1 m thick organic layer. The blade scalp treatment had higher mean temperatures at a depth of 0.1 m compared with the herbicide treatment which retained the organic layer. Daily average temperatures at the 0.1-m depth in late July for the untreated, blade, and herbicide treatments were 10, 13, and 11.5°C, respectively. A small patch scalp treatment at the same site had a slightly lower temperature than the larger blade-scalped area because of the greater shading of the small area.

Macadam (1989b) found that soil temperature increased in response to scalping at a site in the Sub-Boreal Spruce zone (SBSmc2) at 1010 m elevation north of Houston, B.C. (Figure 2.9). Increase in temperature was greatest in the mesic moisture regimes. Because subhygric regimes are wetter than the mesic, the greater heat capacity results in a cooler soil. Scalping removed 5 and 10 cm of organic material from the mesic and subhygric sites, respectively. Removing 15 cm of organic material in a hygric moisture regime did not improve the thermal regime, probably because the exposed soil was as wet as the untreated area. Improved drainage of the soil was required to reduce the heat capacity and increase soil warming (see Section 2.2.4).

There are some disadvantages to using scalping treatments to expose mineral soil. One is that it removes the nutrient pool in the surface organic layer. Another is that the scraping action may close off pores at the soil surface in fine-textured soils. This reduces the rate of infiltration of water and may cause ponding around the seedling. The soil bulk density may also be increased during treatment (Pritchett 1979; Page-Dumroese *et al.* 1986; Childs *et al.* 1990).

2.2.3 Cultivating the soil

Site treatments that cultivate the soil include rototilling, trenching, and mounding. All of them expose the mineral soil and disturb it so that bulk density is reduced. The surface litter or organic layer is often incorporated in the treated area rather than being removed from it, as is the case for scalping treatments. The reduced bulk density may improve drainage and soil aeration, and improve warming by decreasing the volumetric heat capacity of the soil (see Section 3.4). Forming ridges and mounds further improves soil drying and warming. These treatments provide a range of planting spots, depending on the dryness of the environment.

The benefits to the seedlings' aerial environment are also important. These include increased light availability and reduction of vegetation press. Incorporation of the organic layer within the mineral soil keeps most of the soil nutrients within the seedling root zone. The warmer soil temperatures also increase nitrogen mineralization.

Plowing

Plowing the soil exposes and loosens the mineral surface. Plowing may produce small ridges that expose a larger area of soil to the sun, increasing energy input during the day. The influence of exposing the mineral soil and loosening the upper layer is shown in Figure 2.10. The site of this experiment was a flat area northwest of Dawson Creek in the Boreal White and Black Spruce zone (BWBSmw1), at 760 m elevation. The plowing produced small ridges, while the tines on the rake removed more of the vegetation than did the winter shear blade.

Plowing increased warming more than spot scalping in mesic and subhygric regimes of the SBSmc2 site noted earlier (Figure 2.9) (Macadam 1989b). Less warming occurred in subhygric sites. Soil moisture measurements showed the plowed treatments to be drier than the scalped areas.

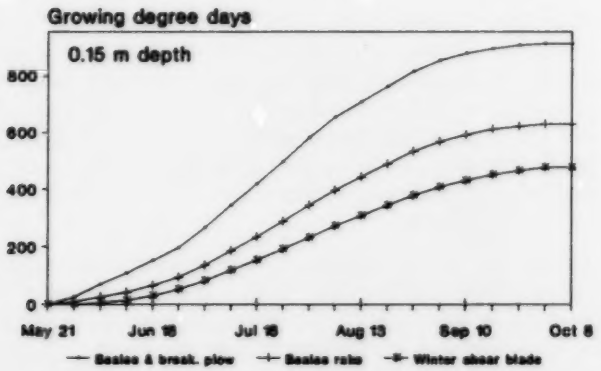

FIGURE 2.10. Accumulated growing degree days (base temperature = 5°C) at the 15 cm depth for the winter shear blade, Beales rake, and Beales rake and breaking plow treatments. The site is in the Boreal White and Black Spruce zone (BWBSmw1) northwest of Dawson Creek, B.C. (T. Letchford, 1986, B.C. Forest Service, unpublished data.)

Disc trenching

Disc trenching can be beneficial on sites where the soil is compact, or where summer drought or frost occur (Beaudry 1989). A variety of planting spots are created. Soil temperature varies with trench orientation (i.e., north/south or east/west) and position in the trench (Table 2.1). South-facing microsites of east/west-oriented trenches are the warmest. Temperatures in the base of the trench are affected by shading from trench walls, with less shading in north/south-oriented trenches.

TABLE 2.1. Cumulative growing degrees (base temperature = 5°C) from the time of planting until the end of the growing season for various site preparation treatments. A threshold temperature of 5°C was used for the growing degree calculation. The site has a clay soil and 5 cm thick organic layer. It is located in the Sub-Boreal Spruce zone (SBSdw3) west of Prince George, B.C. (M. Osberg, 1989, B.C. Forest Service, unpublished data.)

Treatment	Growing degrees
Untreated	1040
Disc trenching	
south aspect	1130
north aspect	950
east/west aspect	945
bottom, E/W-oriented trench	1040
bottom, N/S-oriented trench	1109
Rototilling	1300

Ripping a deep trench in the ground also decreases the soil bulk density and increases its water holding capacity and drainage characteristics, thereby making more water available for seedlings. Summer precipitation tends to run off and collect in trenches. Deep trenches can protect seedlings from radiation frost during the summer. Heat stored in the soil profile during the day is conducted back to the face of the trench and re-radiated toward the seedling at night. This increases the foliage temperature and decreases the risk of frost damage.

Rototilling

This procedure mixes the surface organic layer with the mineral soil to create a berm. Average soil temperature can be increased by up to 4°C compared to that in an untreated area (Figure 2.11). Maximum temperatures were increased by 2-8°C and minimums decreased by 0-2°C. Table 2.1 indicates that rototilling gave the greatest increase in temperature of treatments applied at a warm site in the Sub-Boreal Spruce zone.

Mounding

Cold, wet sites usually require more than exposing and loosening the soil surface layers. Mounds are raised microsites that mimic the raised areas on which natural regeneration often occurs at wet sites. Mounds (and large ridges) have been used in cold or wet environments to create warmer and drier regimes (Draper *et al.* 1985, 1988; McMinn 1985; Page-Dumroese *et al.* 1986; Haeussler 1989; Macadam 1989b; Örlander *et al.* 1990). Mounds consist of a mineral soil capping over either mineral soil, or one or two surface organic layers. They affect soil temperature in a number of ways. Shading vegetation and insulating organic layer are removed,

drainage is improved, and there is a large surface area of exposed mineral soil for heat absorption. Mounds therefore warm up and cool off more rapidly than the surrounding flat ground, although they result in a net warming during the spring and summer compared with other treatments.

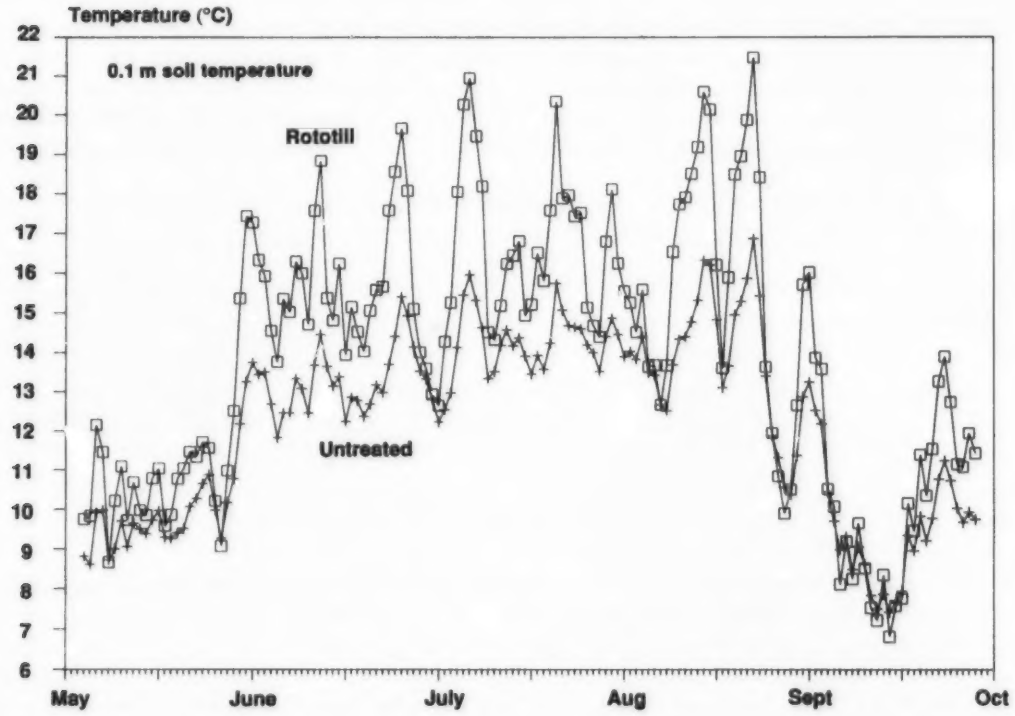

FIGURE 2.11. Daily average temperature at the 0.1 m depth in the berm created by the Madge Rotoclear and in an untreated area. The site has a clay soil with a 5 cm thick organic layer. It is in the Sub-Boreal Spruce zone (SBSdw3) west of Prince George, B.C. (M. Osberg, 1989, B.C. Forest Service, unpublished data.)

Figures 2.8 and 2.9 show the added benefit of mounding over simply exposing the mineral soil by scalping. Temperatures at the 0.1-m depth in the mounded treatment in Figure 2.8 reached an average of 17.5°C in late July; in the control treatment they were only 10.0°C. A herbicide application to the mounded treatment removed any shading from the surrounding vegetation and increased the temperature by an additional 1°C. In May and June, the mounds warmed more rapidly than the other treatments. Mounding also caused larger diurnal temperature fluctuations (Figure 2.12). The height and width of a mound or ridge influences its effectiveness. Large mounds usually warm up more rapidly in the spring but also cool off sooner in the fall (Draper *et al.* 1988).

In dry environments, the increase in surface area for soil drying can put seedlings growing in mounds at greater risk of drought stress during the summer (Örlander *et al.* 1990). On cold, wet sites this is usually not a problem because there is adequate summer precipitation (Draper *et al.*

1985; Macadam 1989b). Mounding creates a favourable aerial environment for the seedlings by reducing vegetation competition. The seedlings receive more light and there is a reduction in vegetation press.

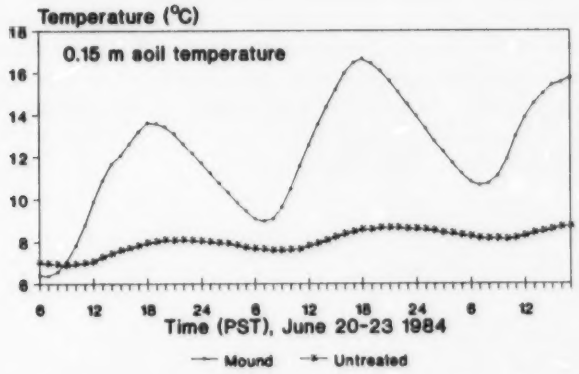

FIGURE 2.12. Variation in soil temperature at the 15 cm depth in a mound and untreated area in the Sub-Boreal Spruce zone (SBSvk) southeast of Prince George, B.C. (See also Figure 2.8.) The untreated area had a dense covering of ladyfern, 0.7 m tall, and a 10 cm thick surface organic layer. (Adapted from Draper *et al.* 1985.)

2.3 Frost Heaving

Frost heaving requires a combination of specific soil temperature, soil moisture, and soil textural conditions (Spittlehouse and Stathers 1990). It is most likely to occur in exposed fine-textured mineral soils that are subjected to repeated freezing and thawing cycles (Örlander *et al.* 1990). Deep planting of seedlings while the soil is warm enough for good root growth to anchor the seedling can minimize this problem.

2.4 Persistence of Treatment Effects

Vegetation regrowth around and within a treatment reduces the effectiveness of the treatment. The reduction of soil temperature by shading can be seen within 2 years in many ecosystems. The influence of the removal of the organic layer remains for a longer period. Mounds described by Draper *et al.* (1985, 1988) were still 2-4°C warmer than untreated areas after 5 years. However, warming the soil in the first 2 years of the seedling's life is probably when the greatest gains would be made. These periods are critical for seedling establishment because they encourage rapid root growth.

3 FACTORS AFFECTING SOIL TEMPERATURE

Soil temperature is affected by site location, atmospheric weather conditions, ground cover, and the physical properties of the soil profile (Pritchett 1979; Spittlehouse and Stathers 1990). The location of the site (latitude, elevation, slope position and gradient, and aspect) influences the receipt of solar energy. Atmospheric weather conditions determine the supply of radiation, heat and water to the ground surface. Vegetation and snow cover modify the effects of atmospheric weather by shading and insulating the soil surface. The thermal and hydraulic properties of the soil profile in turn govern its response to the changing supply of heat and water. The factors which interact to influence the soil thermal regime are summarized in Table 3.1. The difference between heat and temperature is explained in Appendix A and heat flow theory is presented in Appendices B and C.

TABLE 3.1. Summary of factors affecting the soil thermal regime

Category	Factor	Influences
Weather	Solar radiation, air temperature, precipitation, and wind speed	heat transfer into the soil and soil water content
Site	Latitude, elevation, slope, and aspect	solar radiation, air temperature, soil water content, and day length
Surface	Vegetation cover, snow cover, albedo, and surface roughness	solar radiation absorbed
Soil	Soil composition, bulk density, and soil water content	thermal conductivity, volumetric heat capacity, and heat transfer into the soil

3.1 Surface Energy Balance

Solar radiation is the primary source of energy received at the soil surface. Site location influences the amount received; surface characteristics influence the amount reflected and absorbed. Energy is also gained and lost through **longwave radiative heat transfer** between the ground surface and the atmosphere. The net gain of this radiant energy (**net radiation**) is dissipated to the atmosphere and soil as **sensible** and **latent heat**. Sensible heat is heat that can be sensed with a thermometer as a change in temperature. Latent heat is the heat required for the phase change between liquid water and water vapour or ice. Partitioning of the absorbed energy at the surface is described by the energy balance equation (Campbell 1977):

$$Rn = H + LE + G \qquad (1)$$

where Rn is the flux density of net radiation received at the soil surface; H and LE are the flux densities of sensible and latent heat, respectively, which are dissipated into the atmosphere; and G is the heat flow into the soil profile (**soil heat flux**). Atmospheric sensible and latent heat are transferred between the soil surface and the atmosphere by the process of convection. The soil heat flux is transferred by conduction.

Figure 3.1 shows a typical daytime energy balance for a moist soil, where the sensible, latent and soil heat fluxes are all directed away from the surface. The relative amount of each flux depends on whether the soil is bare or vegetated, and the type of vegetation. The partitioning of the net radiation is also strongly influenced by the humidity, temperature and wind speed of the air, the soil water content, and the soil thermal properties.

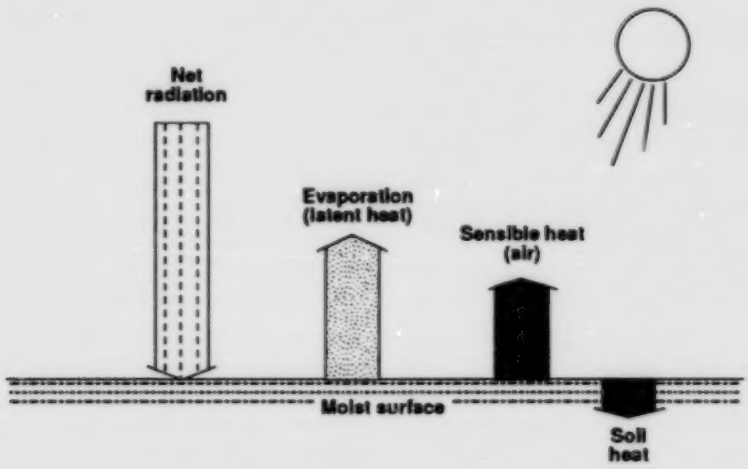

FIGURE 3.1. Daytime energy balance for a bare soil surface.

At night, heat stored in the soil profile during the day is conducted back toward the surface and radiated toward the atmosphere. The net radiation flux is from the surface to the atmosphere. The sensible heat flux, and latent heat flux when dew forms, are directed towards the surface.

A bare soil surface rapidly dries in the top 5-10 cm. This reduces the latent heat flux to a small amount after 5-10 days. Also, the surface temperature rises and, as a result, the sensible heat flux (and, to a small extent, soil heat flux) increase. A vegetated surface can maintain a high transpiration rate for much longer than a bare soil can (Campbell 1977; Spittlehouse and Stathers 1990).

Between 5 and 40% of the net radiation is transferred into the soil profile during the day. The diurnal variation of the soil heat flux largely depends on how the surface temperature gradient, the soil thermal conductivity, and near-surface soil and air temperatures change during the day. Vegetated (shaded) surfaces have a much lower soil heat flux than a bare soil surface, and are consequently much cooler.

3.2 Surface Radiation Balance

Net radiation is the sum of the absorbed solar and longwave radiation fluxes at the surface. Direct and diffuse solar radiation (K_{down}) are received from the sky and a fraction of this is reflected back from the surface (K_{up}). Longwave (or thermal) radiation is emitted by the atmosphere (L_{down}) and the surface (L_{up}) as a function of their respective absolute temperatures. The net radiation at the ground surface is given by (Campbell 1977):

$$Rn = (K_{down} - K_{up}) + (L_{down} - L_{up}) \qquad (2)$$

Figure 3.2 shows a typical daytime radiation balance for a bare soil surface where there is a gain in energy. At night, net radiation consists of only the longwave component. Since the effective sky temperature is much lower than the ground surface, there is a net loss of radiant energy from the soil surface toward the sky. Radiative cooling is thus the main process by which the heat stored in the soil

profile during the day is lost at night. In the fall and winter, or on cloudy days during the summer, there also tends to be a daily net loss of heat from the soil profile.

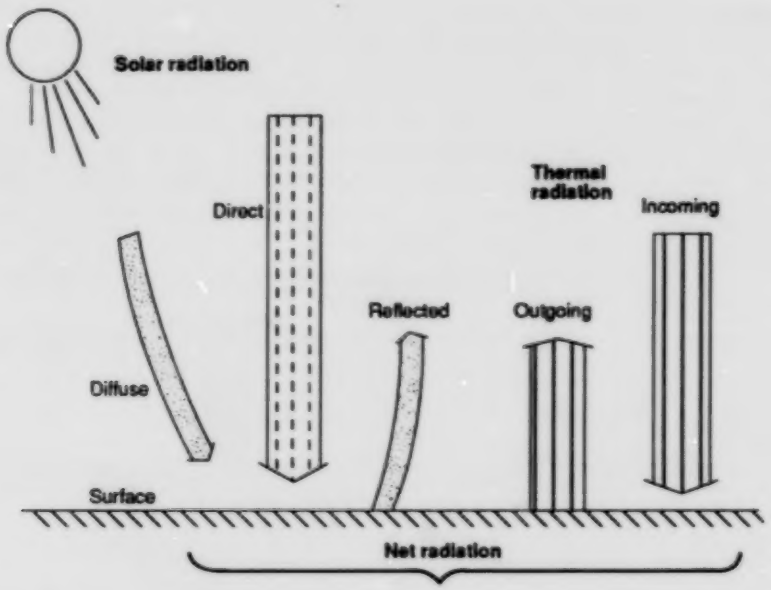

FIGURE 3.2. Daytime radiation balance for a bare soil surface.

The **albedo** (or **reflectivity**) of the surface (= K_{up}/K_{down}) influences the amount of solar radiation available to heat the surface during the day. The albedo of most dry mineral soil surfaces and vegetation canopies ranges from 0.10 to 0.30. Wet mineral or organic surfaces have albedos that range from 0.05 to 0.15. A freshly burned surface can have an albedo of 0.02. As the soil surface dries, the amount of solar radiation reflected from the surface increases and the amount absorbed decreases.

The amount of longwave radiation emitted by the soil surface and the vegetation depends on the fourth power of its absolute temperature and its **emissivity**. The emissivity of soils and vegetation is not as variable as the albedo, and ranges from about 0.90 for dry soils to 0.96 for wet soil and vegetation. The effective emissivity of a clear sky is about 0.75.

3.3 Site-related Factors Affecting Soil Temperature

A large proportion of the variation in soil temperature over the landscape can be attributed to the effects of latitude, elevation, slope, and aspect on the daily and seasonal duration and intensity of solar radiation. These factors also indirectly influence the soil thermal regime through their effect on site drainage, vegetation or snow cover, and weather conditions.

Weather conditions directly influence soil temperature. Under clear sky conditions, ground surface temperatures are warmer during the day because of the increased solar irradiance, and colder at night because of the increased longwave emittance from the surface. **Cloud** reduces the solar irradiance during the day and increases the longwave emittance from the sky at night. This results in a decrease in the maximum and increase in the minimum temperature of the soil compared to those temperatures under a clear sky. **Rainfall** has a small direct effect on near-surface soil temperatures. The influence of snow is discussed below. Wind reduces the surface temperature during the day by causing more of the energy absorbed at the ground surface to be dissipated to the atmosphere rather than being conducted into the soil.

Latitude influences the soil thermal regime through its effect on day length. As the number of daylight hours increases, the change in soil profile heat storage increases because more heat is transferred into the profile during the longer day and less heat is radiated back to the atmosphere during the shorter night. At higher latitudes, near-surface soil temperatures can be similar to those which occur at more southerly locations during the summer because the increased day length tends to compensate for the shorter summer season. Soil temperatures can vary to a depth of 10 m during the course of a year, and variation is much greater at higher latitudes because of the greater variation in solar irradiance.

The average air temperature decreases and precipitation increases with increasing **elevation** (Oke 1978). Cooler, wetter conditions generally translate into cooler soil temperature regimes.

The **proximity of the site to the ocean or a large lake** influences the soil thermal regime. Large water bodies act as heat sinks and water vapour sources, which tend to reduce air temperature extremes and increase precipitation (Oke 1978).

The effect of **slope** and **aspect** on the diurnal and annual receipt of solar energy is shown in Figure 3.3. During a year, steep south-facing slopes can receive up to three times as much clear sky solar radiation as north-facing slopes (Buffo *et al.* 1972). The effect of slope and aspect on solar irradiance is greatest in the early spring, late fall, and winter when the solar elevation angle is lowest. As a result, south-facing slopes tend to warm up more rapidly than other aspects in the early spring, particularly at higher latitudes and on steeper slopes.

Vegetative cover affects soil temperature by absorbing solar radiation and shading the soil surface. It also reduces the rate of longwave radiative cooling of the soil surface at night. The temperature of the vegetation canopy is much higher than the effective clear sky temperature (about -20°C at night). As a result, the canopy absorbs and re-radiates some of the radiation emitted by the ground surface, and thereby reduces the rate of ground surface cooling.

The effect of vegetation is illustrated in Figure 3.4. This shows a comparison of soil temperatures in a forest clearcut and adjacent mature western hemlock and Douglas-fir forest during a clear, hot summer day. Although soil temperatures below the 0.5-m depth were similar, the diurnal soil temperature variation nearer the surface was much greater in the clearcut. Maximum surface soil temperatures were higher than 50°C in the clearcut, but were only 16°C in the forest. Generally, under a dense forest canopy, the near-surface soil temperatures are similar to air temperatures during the day and at night.

Snow cover can benefit soil temperature, particularly in cold climates. Snow is a poor heat conductor, and a snow pack acts as an insulating layer that reduces the rate of heat loss from the soil. This tends to keep near-surface soil temperatures near 0°C (Figure 3.5), which reduces both the depth of frost penetration in the winter and the amount of heat required to warm the soil profile again in the spring. The depth to which soil freezing occurs in winter depends on the duration and severity of cold atmospheric conditions and the depth and duration of snow cover. Cold weather in the late fall before the snow falls, or intermittent snowfall and melting during the winter, can result in low soil temperatures. Deep snow packs that do not melt until early summer have a negative effect on seedling root zone temperature by delaying soil warming.

3.4 Soil-related Factors Affecting Soil Temperature

Heat storage, heat transfer, and the temperature within the soil profile are determined by the thermal conductivity and volumetric heat capacity of the soil. These thermal properties can vary considerably with the composition (texture, structure, organic matter and coarse fragment content), bulk density, and water content of the soil profile (Appendix B). Thus, under the same environmental conditions there can be substantial variation in the thermal response of different soils. Similarly, thermal response can vary with depth in the soil profile (Campbell 1977, 1985; Hanks and Ashcroft 1980).

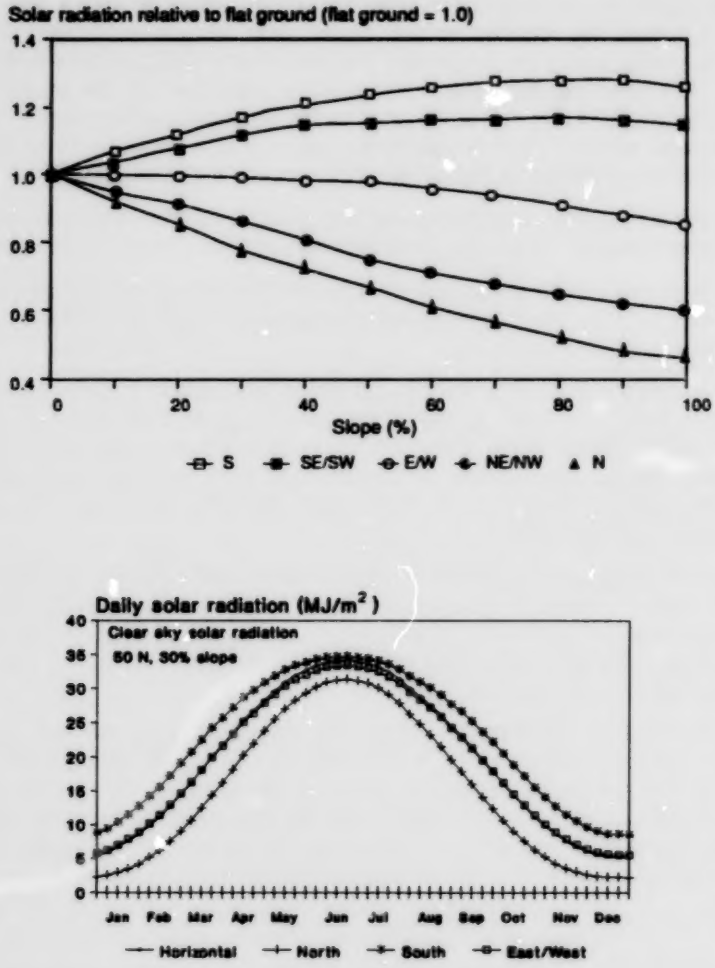

FIGURE 3.3. Clear sky solar irradiance on various slopes and aspects at 50°N latitude. The upper graph shows the annual amount received on selected aspects and slopes. The lower graph shows the annual variation in the daily total for a 30% slope (20°) on selected aspects.

The **volumetric heat capacity** is the amount of heat required to cause a temperature change of 1°C in a volume of soil. The larger its value, the smaller the change in the soil temperature for the same amount of heat. The heat capacity and the amount of heat stored in the soil at a depth determine the soil temperature (Appendix A). The **thermal conductivity** of the soil determines the rate of heat transfer through the soil under the conditions of a given **temperature gradient**. Heat transfer through the soil profile is outlined in greater detail in Appendix C.

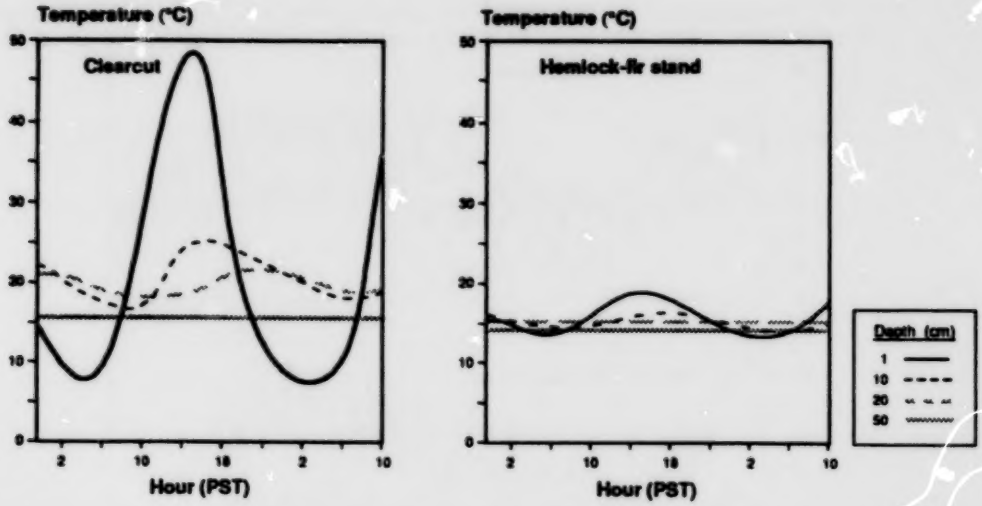

FIGURE 3.4. Diurnal courses of soil temperature during a hot summer day in a southwest-facing forest clearcut and adjacent mature western hemlock and Douglas-fir forest on a south-facing slope, 1150 m, Coastal Western Hemlock zone (CWHmm2), east of Port Alberni, B.C. (Adapted from Stathers 1983.)

Soil is a three-phase medium composed of **solids** and of voids filled with **water** and **air**. As the soil water content changes, the air and water fractions displace each other. Air and water have different thermal conductivities and volumetric heat capacities. Consequently, the soil moisture regime has a large influence on the soil thermal regime. Thermal conductivity and volumetric heat capacity increase as the soil water content increases. The thermal properties of various soil constituents relative to those for dry mineral soil are given in Table 3.2.

The **thermal diffusivity** is the ratio of the thermal conductivity to the volumetric heat capacity. The thermal diffusivity determines how rapidly temperature changes are transmitted through the profile; thermal conductivity determines the rate of heat conduction through the profile. Thermal diffusivity varies with soil composition and water content (Figure 3.6).

Dry mineral soils have a low thermal diffusivity. It increases rapidly as the volumetric soil water content increases to about 10 or 15% and then declines again as the soil water content approaches saturation. As a result, temperature changes are transmitted relatively slowly through dry, or wet, mineral soils. Also, a wet soil tends to have a lower surface temperature than a drier one, since more energy is lost to water evaporation. Consequently, wet soils warm more slowly and are generally colder than drier soils in the same area. Fine-textured soils are often wetter, and therefore cooler, than coarser-textured soils. Coarse-textured soils tend to create a surface mulch more readily than fine-textured soils, and this tends to decrease both evaporative cooling of the surface and heat flow into the profile.

Surface organic layers have a high water holding capacity and a low thermal diffusivity at all water contents. As a result, a thick wet organic layer, which has a high volumetric heat capacity, is an effective insulator which reduces heat penetration into underlying mineral soil. Surface organic layers also act as good insulators because they have a low thermal conductivity when either wet or dry. Under clear sky conditions, dry organic surfaces usually get much warmer during the day and colder at night than do mineral soils (Figures 2.5, 2.6, and 3.7). This occurs because the low thermal conductivity of organic matter reduces heat transfer into the profile and the low volumetric heat capacity causes a relatively large change in temperature for a small change in heat storage.

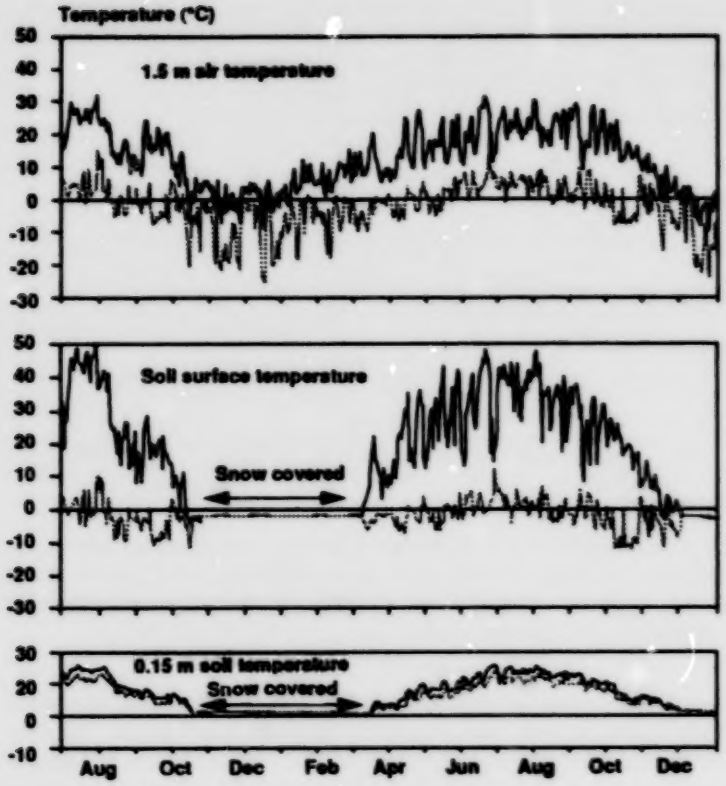

FIGURE 3.5. Annual variation in maximum and minimum air and soil (surface and 15 cm) temperatures in a clearcut in the Montane Spruce zone (MSdm2), 1130 m, flat, gravelly sandy loam, south of Gold Bridge, B.C. There was an intermittent layer of blackened organic material (up to 2 cm thick) and minimal vegetation. (Adapted from Spittlehouse and Stathers 1990.)

TABLE 3.2. Thermal properties of soil constituents relative to those for a dry mineral soil

Material	Volumetric heat capacity	Thermal conductivity
Dry mineral soil	1.0	1.0
Dry organic soil	0.3	0.2
Air	0.001	0.1
Water	4.0	2.0

Mounding, ridging, and **disc trenching treatments** influence the thermal regime of the seedling root zone (Section 2.2.3) in a number of ways. The raised soil tends to drain better than a flat surface and thus has a lower heat capacity. The sloping sides and relatively large surface have a different radiation balance than the flat surface, absorbing more energy during the day and losing more at night.

Consequently, these treatments show a higher mean temperature during the spring and summer (Figures 2.8 and 2.9, Table 2.1) and a wider fluctuation during the day in the raised portion (Figure 2.12).

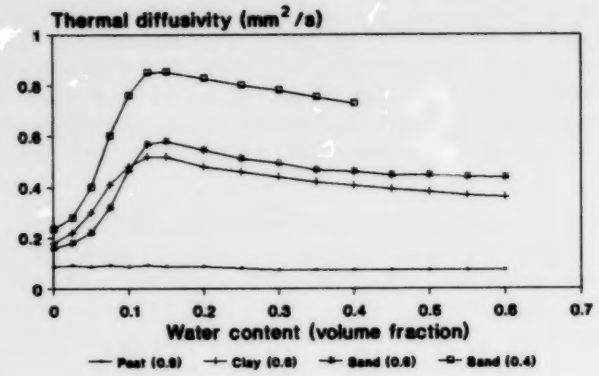

FIGURE 3.6. Effect of volumetric soil water content on thermal diffusivity of peat, clay, and sand. Numbers adjacent to each texture indicate the volume fraction of solids. (Adapted from Campbell 1977.)

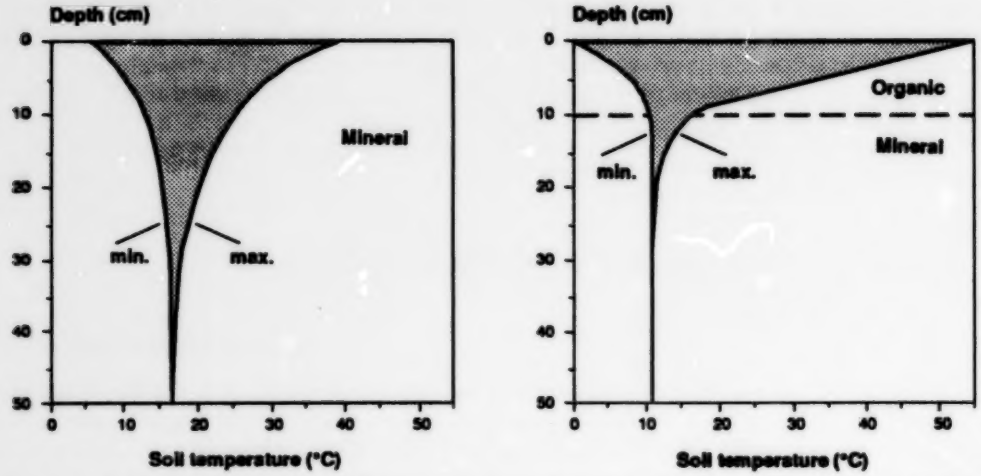

FIGURE 3.7. Daily maximum and minimum soil temperature with depth under the same environmental conditions in a mineral soil and a mineral soil covered with 10 cm of organic material. (Adapted from Cochran 1969.)

3.5 Soil Temperature Dynamics

Soil temperature varies continuously in response to changes in energy receipt and partitioning at the soil surface. Consequently, there are distinct **diurnal** and **annual soil temperature cycles** (Figures 2.12, 3.4, and 3.5), just as there are distinct periodic cycles of solar radiation. Over a day, heat is conducted from the soil surface to a depth of about 0.5 m (Figures 2.6 and 3.4), depending on the soil thermal properties and the amount of energy absorbed by the ground surface. Soil temperature changes with depth as heat is redistributed within the profile. Heat is conducted down gradients of temperature, that is, from warmer to cooler regions. Consequently, as the surface absorbs solar radiation and warms

during the daytime, heat is slowly conducted into the cooler underlying soil. At night, the ground surface cools as it emits longwave radiation to the atmosphere, and most of the heat which was stored in the profile during the day is conducted back toward the cooler surface.

The variation in soil temperature is greatest at the soil surface and decreases rapidly with increasing depth. Heat is conducted relatively slowly through soil. It must accumulate until a temperature gradient is generated before it can be conducted deeper into the profile. The slow rate of heat transfer through the profile also causes temperatures at depth to lag increasingly behind changes in surface temperature. The combined effects, on a daily basis, of the decreasing temperature amplitude and increasing temperature phase lag with depth is shown in Figure 3.4. The soil profile is rarely isothermal (i.e., at the same temperature at all depths).

The annual solar cycle produces annual variations in soil temperature to a depth of about 10 m. The slow transfer of heat also results in a temperature lag with depth on an annual basis (Figure 3.8). During the spring and early summer when the soil near the surface is warming, deeper soil (>1 m) is still cooling as the winter portion of the cooling wave penetrates into the profile. The soil temperature at the 1- to 2-m depth typically lags behind near-surface temperatures by about 3 months. The minimum and maximum temperatures at these depths occur in March and September, respectively.

The limited data available suggest that forest soils in the province usually do not freeze to great depths during the winter. There may be some initial freezing in the top 0.2 m of the profile if the snow pack is late in developing or is lost during winter. However, the snow pack insulates the soil from cold winter air temperatures and reduces the rate of heat loss from the profile. The soil may freeze down to about the 0.5- to 1.0-m depth at sites where there is very little snow cover and continuously cold winter weather. The diurnal temperature variation in the soil profile is also negligible during the winter even though atmospheric temperatures are variable (Figure 3.5).

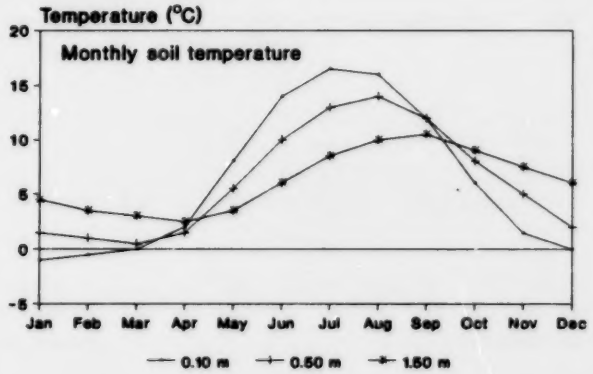

FIGURE 3.8. Monthly average soil temperatures at Prince George, B.C., showing the phase lag in soil temperature with depth.

Mounds and ridges show the above cycles and lags, but have a different temperature distribution in the upper part of the profile. They have a greater daily and annual temperature fluctuation than a flat surface; they warm sooner in the spring and cool faster in the fall; and they have a higher mean summer temperature (Draper *et al.* 1985, 1988; Macadam 1989b; Örlander *et al.* 1990). The maximum temperatures in a mound on a sunny July day are shown Figure 3.9. This mound was conical with a flat top. The time of maximum soil temperature near the surface varied as the sun moved around the mound. Soil temperatures between depths of 5-20 cm remained above 20°C for more than 10 hours. Minimum temperatures in the mound were between 9 and 12°C.

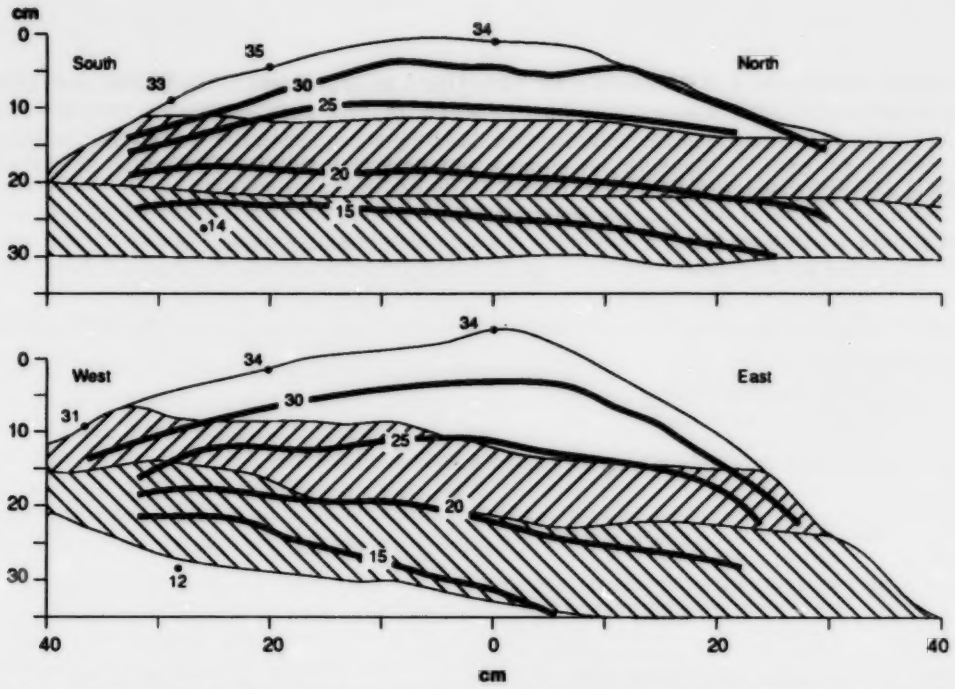

FIGURE 3.9. South-north and west-east slices through a mound showing isotherms of maximum soil temperature (°C) on a sunny July day. Hatching indicates organic layers. The location is the Sub-Boreal Spruce zone (SBSvk), silty-loam soil, northeast-facing slope, southeast of Prince George, B.C. (See also Figures 2.8 and 2.12.) (D. Spittlehouse, 1986, B.C. Ministry of Forests, unpublished data.)

4 MEASURING SOIL TEMPERATURE

4.1 Soil Temperature Sensors

A variety of sensors are available to measure soil temperature (Table 4.1). They can be broadly classified as electrical, mechanical, or chemical, according to how the transducer senses thermal energy and converts it into a signal which is related to the temperature. The best type of sensor for any particular application depends on a number of requirements. Some of the characteristics which should be considered include the accuracy, temperature range, response time (time constant), ruggedness, reliability, cost, ease of installation, and compatibility with measurement and recording devices (Unwin 1980).

One-point temperature measurements for soil survey or planting window assessments can be determined with adequate accuracy using metal-sheathed mercury-in-glass or bimetallic strip thermometers. These instruments are relatively inexpensive and durable enough for field use. Thermistor or thermocouple thermometers are also available for this application, but require an electronic measuring and read-out device. A data logger can be used to measure and record temperatures continuously. These instruments should be considered for studies requiring multiple measurements over long periods of time, particularly for studies at remote locations (Spittlehouse 1989).

A small sensor is required when temperatures change rapidly (i.e., >1°C per minute). Sensors should have a low heat capacity and high thermal conductivity. The sensor should not be so large or have such different thermal properties that it influences the surrounding soil temperature. This is only a concern within the top 2 cm of the soil surface. Deeper in a soil profile the temperature changes relatively slowly.

Platinum resistance thermometers are accurate, but expensive. Thermistor and thermocouple thermometers are inexpensive and can be configured to meet almost any measurement requirement. Chemical sensors are also relatively inexpensive, but they are not particularly accurate. Infra-red sensing thermometers are used to measure the surface temperature remotely. Calibration of soil temperature sensors is described in Appendix D.

TABLE 4.1. Sensors used for measuring soil temperature (Unwin 1980; Omega 1982)

Mechanical sensors
Liquid-in-glass thermometers
Liquid-in-metal thermometers
Gas thermometers
Deformation thermometers (bimetallic strips)
Electrical sensors
Self-generating thermoelectric thermometers
thermocouples
Non-self-generating resistive thermometers
metallic wire resistance thermometers
thermistors
diodes
integrated circuit devices
Infra-red thermometers
Chemical sensors
Reaction kinetics of chemical solutions
Temperature coefficients of electrolytic solutions
Melting points of compounds
Heat sensitive films or crystals which change colour

4.2 Location and Installation of Soil Thermometers

Temperature measurements should be taken at specified depths and times so that inter-site comparisons can be made. The FAO/UNESCO Committee on Soil Climate has recommended that temperatures be measured at the 0.01, 0.05, 0.10, 0.20, 0.50, 1.00, 1.50 and 3.00 m depths. In many studies of forest climate it is not necessary to measure the temperature at all of these depths. Measurements are usually concentrated in the plant root zone. Surface temperature measurements can also be important in some applications. The temperature at the 0.5-m depth is used for soil climate classification. It is recommended as one of the higher priority measurement depths if inter-site comparisons are of interest.

Physical factors such as vegetation cover, soil thermal properties, and soil water content significantly influence soil profile temperatures. For this reason, the measurement site must be carefully selected so that the measurements are representative of the conditions meant to be studied. Where there is considerable spatial variability in surface cover, such as in a forest clearcut, it is necessary to replicate measurements, particularly in the top 0.25 m of the profile. This can be done by either using more sensors or by connecting a number of sensors together in series or parallel to obtain a spatially representative sample. Five sensors are recommended as a minimum. At standard climatological stations, soil temperatures are usually measured either under uniform grass-covered plots where the grass is maintained 5-10 cm long, or under a bare soil surface so that standard sites are comparable in all climatic regions.

Temperature sensors must be installed with minimum disturbance of the soil profile. Altering the soil composition or bulk density can change the surface energy and water balances, which in turn can

adversely influence the heat and moisture flow fields within the profile. This can cause the temperature distribution in the instrumented profile to differ from the surrounding area. A brief description of the procedure for installing sensors into the profile for long-term measurements is given in Appendix E.

Care must be taken when manual measurements are made with a portable thermometer. There must be good contact between the thermometer and the surrounding soil, and time should be allowed for the sensor to equilibrate with the soil before the temperature is recorded. If measurements are being made in a soil pit, the shaded face of the pit should be used to minimize radiative heating of the sensor.

4.3 Sampling Frequency

The sampling frequency is determined by the objectives of the study and the equipment used. If a data logger is used to measure temperature (Spittlehouse 1989), the sampling frequency is usually set to between 10 seconds and 1 minute for sensors between 0 and 30 cm deep. This time may be increased if data logger power consumption is a concern. These measurements are usually averaged over periods of either an hour or a day, depending on the study requirements. Daily maximum and minimum temperatures are often recorded to accompany daily average values. Thus, the temperature variation at a particular location can be characterized by its daily mean value and range.

Single-point manual measurements can be used to characterize the thermal regime. However, because temperature varies with time and depth (Figure 3.4), these must be recorded when the measurements are taken to enable interpretation of the data. Point measurements must be taken between 1200 and 1500 hours to provide the maximum surface temperature, and just before sunrise to provide the minimum surface temperature. Because the temperature phase lag increases with depth, early afternoon measurements provide an estimate of the minimum temperature at about the 0.2-m depth and the daily average temperature at the 0.10- to 0.15-m depth (see Figure 3.4). Temperatures below the 0.4-m depth can be measured at any time because they change by less than 0.2°C per day. Weather conditions should be noted when measurements are taken to help interpretation of near-surface data.

A summary of the types of measurement errors and how to avoid them can be found in Gill and Hexter (1972), Fritschen and Gay (1979), Unwin (1980), and Spittlehouse (1989).

4.4 Computer Simulation of the Effect of Site Preparation Treatments on the Soil Thermal Regime

Computer models can be used to assess how site preparation treatments can affect the soil thermal regime. One of the advantages of modelling treatment effects is that most kinds of site preparation treatments can be simulated under a variety of growing season weather conditions. Modelling also allows a cost-benefit analysis of additional treatments. Furthermore, since the goals of operational treatments are not always met, a model can be used to determine the effect of an incomplete treatment.

The soil temperature model used here is based on the physics of heat, momentum, and radiation transfer (Stathers *et al.* 1985, 1988). The data required to run the model include solar radiation, air temperature, relative humidity, wind speed, precipitation on an hourly or daily basis throughout the growing season, and soil thermal and hydrologic characteristics. The model determines the heat and moisture transfer and storage within the soil profile resulting from the changes in the input weather conditions.

An example of the modelled effects of site preparation on growing season soil profile temperatures is given in Figure 4.1. The model was set to simulate a high elevation clearcut having a dense vegetation canopy. The soil was a compact glacial till overlain with 0.15 m of organic material. The model was run for an untreated, densely vegetated area, for a herbicide treatment which killed the vegetation but did not remove any of the organic layer, for a prescribed burn which left 5 cm of blackened organic material, and for a scalped area where all of the organic material was removed. The growing season was characterized by a cool, wet June and July and a dry, warm period in August. The simulation was run from the time the snow pack melted in early May into late September.

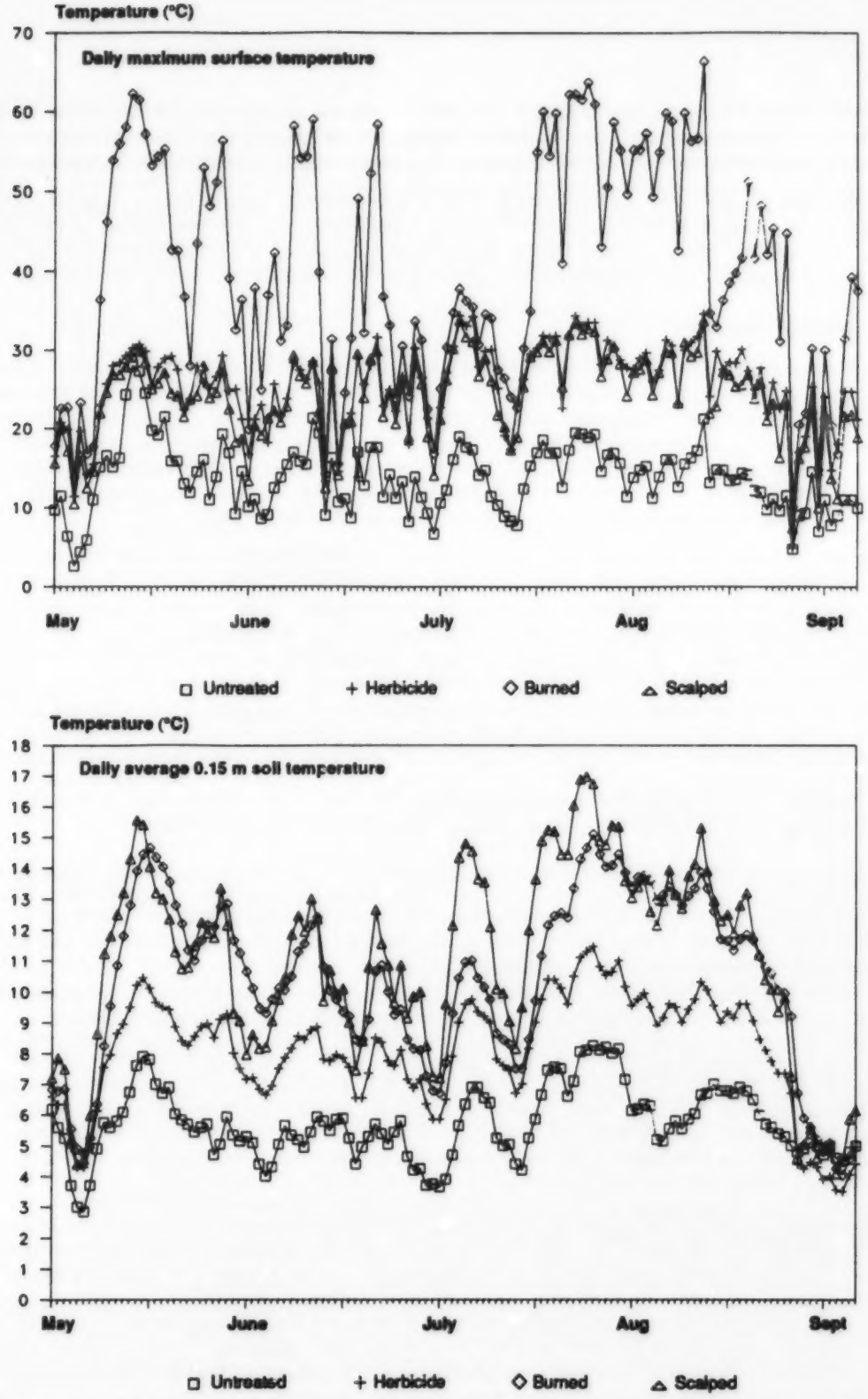

FIGURE 4.1. Computer simulation of the effects of site preparation on daily maximum surface soil temperature and daily average 15 cm soil temperature over a growing season.

The results from these simulations show many similarities to those found in the comparison of treatments shown in Figure 2.7. The untreated brushy area had the lowest root zone soil temperatures, ranging from 4-7°C during the growing season. The herbicide treatment increased root zone temperatures by 2-4°C, but the organic layer continued to act as an insulator. Burning increased root zone temperatures even more. Scalping down to mineral soil had the greatest benefit, increasing root zone temperatures by as much as 10°C. Extremely high surface temperatures occurred in the burned area, resulting from the effect of the low albedo and thermal properties of the organic layer on solar radiation absorption and heat transfer.

5 INTERPRETING SOIL TEMPERATURE MEASUREMENTS

5.1 Soil Thermal Regime Classification

The characterization of soil thermal regimes for use in climate studies in North America is based on soil temperature during periods of biological significance. The simplest definitions use growing and dormant season soil temperatures at the depth of interest (Baier and Mack 1973):

Growing Season	> 5°C
mild periods	> 5°C
warm periods	>15°C
Dormant Season	< 5°C
cool periods	> 0°C
frozen periods	< 0°C

This division of the year into the dormant and growing seasons has been based on observations for agricultural crops, which tend to grow only when soil temperatures exceed 5°C. Some coniferous species, however, are adapted to cold climates and can grow slowly at root zone temperatures as low as 2°C. Forests may consequently have a longer growing season than that indicated by the agriculture-based definitions. Indices determined from temperatures at the 0.5-m depth are also used to classify thermal regimes (Table 5.1), e.g., Baier and Mack (1973), Vold and Marsh (1981).

TABLE 5.1. Indices used to classify soil thermal regimes based on the temperature at the 0.5-m depth

Mean annual soil temperature (MAST)	°C
Mean summer soil temperature (MSST)	°C
Length of growing season	days > 5°C
Growing season degree days	> 5°C
Length of thermal period	days >15°C
Thermal period degree days	>15°C
Length of dormant season	days < 5°C
Mean dormant season temperature	°C

Seven major soil temperature classes have been defined on the basis of these indices by the FAO/UNESCO Soil Climate Committee. The classes express an increasing degree of warmth of soil climate and are named arctic, subarctic, cryoboreal, boreal, mesic, thermic, and hyperthermic. In Canada, 30.0, 25.7, 34.4, 8.6, and 1.3% of the land area, respectively, is found in the first five land classes. The thermic and hyperthermic soil temperature regimes do not occur in Canada. Characteristics of each of these classes (using definitions established for the FAO/UNESCO Soil Climate Map of North America) are given in Table 5.2.

These soil temperature classes were developed for continental climates, characterized by wide diurnal and seasonal fluctuations in temperature. Maritime or mountainous regions may experience longer or shorter growing seasons and have significantly fewer growing degree days.

TABLE 5.2. Soil temperature classes based on the soil temperature at the 0.5-m depth. FAO/UNESCO soil temperature classification nomenclature used. The Thermic and Hyperthermic regimes (mean annual soil temperatures of 15<22°C and ≥22°C, respectively) are not shown. (Modified from Luttmerding *et al.* 1990.)

FAO/ UNESCO	Soil temperature classes	Description	Differentiating characteristics					
			MAST[a]	MSST[b]	Growing season (Days >5°C)	Degree-days (>5°)	Warm thermal period (Days>15°C)	Degree-days (15°C)
Arctic	Extremely cold	Continuous permafrost usually occurs within 1 m of the surface	<-7°C	<5°C	<15	0	0	0
Subarctic	Very cold	Soils frozen during the dormant season; discontinuous permafrost may occur	-7 to 2°C	5 to 8°C	<120	<550	0	0
Cryoboreal	Cold	Soils usually frozen in some part of the control section in part of the growing season, no permafrost	2 to 8°C	8 to 15°C	140-220	550-1250	0-50	<30
Boreal	Cool	Soils may or may not be frozen in part of the control section for a short part of the dormant season	5 to 8°C	15 to 18°C	170-220	1250-1700	>60	30-220
Mesic	Mild	Soils are rarely frozen during the dormant season	8 to 15°C	15 to 22°C	200-365	1700-2800	90-180	170-670

a MAST: mean annual soil temperature.
b MSST: mean summer soil temperature.

5.2 Soil Temperature and Tree Growth

The concept of cardinal temperature limits is useful for interpreting soil temperature environments for plant growth. Plants grow well within a relatively narrow range of temperatures because of the effect of temperature on physiological processes such as photosynthesis, transpiration, respiration, and water or nutrient uptake by roots (Gates 1980; Jones 1983). Growth does not occur below a critical minimum temperature or threshold temperature because these physiological processes are temperature limited. For most conifer seedlings this temperature has been observed to be between 2 and 5°C in the root zone, depending on the species (see Figures 1.1 and 1.2; and Heninger and White 1974).

Soil temperature affects uptake of water by seedlings (Figure 1.2). Low soil temperatures can impose a physiological drought on seedlings since the root resistance to water uptake increases as soil temperature decreases (Örlander *et al.* 1990). Seedlings planted on sites which typically have a short growing season and period of drought during the summer will often show symptoms of moisture stress because the low soil temperatures during the spring inhibit root growth and thereby reduce the volume of soil from which roots can extract water.

Both root and shoot growth rates increase as soil and air temperatures increase and become less limiting. Maximum growth rates occur at optimal soil and air temperatures ranging from 15 to 30°C. Optimal temperatures vary for different species (Lavender and Overton 1972; Heninger and White 1974). For Douglas-fir, maximum growth rates have been reported to occur at root zone temperatures of 18-24°C; white spruce at 19°C; jack pine at 27°C; Monterey pine at 15°C; California redwood at 19°C; and paper birch at 19-31°C.

Soil temperatures above 30-50°C for extended periods can cause indirect temperature stress (e.g., Seidel 1986). This reduces the potential growth rate by affecting photosynthesis and respiration. At temperatures above 50-60°C, direct temperature stress for brief periods can cause bark lesions, scald spots, and desiccation of the cambium (Seidel 1986). This damage is usually evident at the root collar near the soil surface, where the highest temperatures are most likely to occur. Cambial damage results in reduced water flow through the stem and this can cause moisture stress in the seedling. Excessively high temperatures also cause cell membrane damage, protein and enzyme denaturation and the accumulation of toxic nitrogenous compounds which can also cause seedling mortality.

High soil and air temperature acclimation is not as well developed in tree species as is low temperature acclimation. Winter air temperatures as low as -40 to -50°C do not generally damage most western coniferous species if they are in a hardened, dormant condition. However, if temperatures of about -3°C occur during the active growing season, foliage or succulent stem tissue can be frost-damaged. Roots are even more susceptible to frost damage than shoots (Sakai and Larcher 1987). Frost injury is caused by the intracellular freezing of cell contents or the dehydration of intracellular protoplasm as intercellular water freezes.

The coldest temperature which can be survived is thought to determine the northern extent of distribution of many tree species. Tree species distribution is probably limited by the occasional severely cold winter temperatures rather than the average winter minimum temperature. Frost damage during the active growing season is probably also an important determinant of the geographic range of tree species.

6 SUMMARY

The soil thermal regime is modified by almost every forest management operation. Clearcutting, thinning, spacing, pruning, and site preparation treatments alter the solar and longwave radiation regimes and consequently affect the amount of energy received at the ground surface. These treatments also influence snow accumulation and the rate of snow melt, and so affect frost penetration and winter minimum soil temperatures.

Forest harvesting can influence the thermal and hydraulic properties of the profile by changing the soil composition and bulk density. Heavy machinery can scarify or mound the surface horizons and compact the lower horizons. Prescribed burning blackens the surface and reduces the depth of the litter layer, thereby modifying the thermal and radiative properties of the surface. Road building can significantly change local drainage patterns, which in turn can influence the soil water content. Harvesting changes the hydrologic characteristics of the soil profile as well as the growing season water balance. These changes in the soil water regime influence the soil thermal regime through their subsequent effects on the soil thermal properties and the evaporation rate from the surface.

Site preparation treatments can be used deliberately to alter the thermal regime of a site. However, there are often trade-offs between gains in soil temperature and losses in other resources. Treatments aimed at altering one resource invariably influence other resources.

Understanding how forestry operations influence soil climate should help the forester in choosing operational procedures which optimize the growth potential of the site.

7 GLOSSARY OF TERMS

ABSORPTIVITY: the fraction of incident radiation of a given wavelength that is absorbed by a surface.

ALBEDO: the ratio of the amount of solar radiation reflected by a surface to the amount incident on the surface.

ABSOLUTE TEMPERATURE: the temperature of a body or surface referenced to absolute zero and expressed in units of Kelvins (K). (0°C = 273.15 K)

AMPLITUDE: (maximum temperature - minimum temperature) / 2.

BLACK BODY: a perfect longwave emitter which emits radiation in proportion to the fourth power of its absolute temperature.

CONDUCTION: the exchange of heat or mass in space by apparently random motions on a molecular scale.

CONVECTION: the exchange of heat or mass in space by random movement of a mass of fluid (e.g., air or water).

DAMPING DEPTH: an index of the depth of penetration of surface temperature changes in a given period of time. It depends on the soil thermal conductivity and volumetric heat capacity of the soil.

EMISSIVITY: relates to the loss of longwave or thermal energy from a surface. It is the ratio of the amount emitted by the surface to that emitted from a perfect emitter (a black body) at the same temperature.

EVAPORATION: the process by which liquid water is transformed into water vapour.

FLUX: rate of flow of some quantity (e.g., Joule per second [= Watt]).

FLUX DENSITY: the flux through a unit surface area (e.g., Watt per square metre).

GRADIENT: variation of an environmental parameter such as air temperature with distance, height or depth.

GROWING DEGREES: the number of degrees the daily mean temperature is above a set value (often 5°C). For example, a day with a mean temperature of 10°C has 5 growing degrees. Growing degrees are usually accumulated over a number of days.

IRRADIANCE: the amount (flux density) of solar or net radiation received by a surface.

LATENT HEAT: the heat absorbed or released per unit mass of water during a change in phase. For example, the latent heat of vapourization refers to the change from liquid to vapour. That from liquid to solid is the latent heat of fusion.

LATENT HEAT FLUX: the proportion of the surface energy balance in which energy is dissipated through evaporation or condensation of water.

LONGWAVE RADIATION: thermal radiation (4-80 micrometre wavelengths) emitted from a surface as a function of the absolute temperature and emissivity of the surface.

NET RADIATION: the sum of the absorbed shortwave and longwave radiation fluxes at the surface.

RADIATION: electromagnetic energy. (In this manual we are concerned with the solar and longwave components.)

REFLECTIVITY: the fraction of incident radiation at a given wavelength reflected by a surface.

RELATIVE HUMIDITY: the relative saturation of the air. That is, the ratio of the vapour density of the air to the saturation vapour density of air at that temperature.

SENSIBLE HEAT: heat which can be sensed with a thermometer.

SENSIBLE HEAT FLUX: the proportion of the surface energy balance in which energy is dissipated into or received from the atmosphere as sensible heat.

SOIL HEAT FLUX: the proportion of the surface energy balance in which energy is transferred into or out of the soil.

SOIL HYDRAULIC PROPERTIES: pertains to the soil hydraulic conductivity and soil water retention characteristic which determine the transfer and storage of water in the soil profile.

SOLAR RADIATION: energy radiated from the sun. Over 90% of the energy is between wavelengths of 0.3-4 micrometres.

SPECIFIC HEAT: the heat capacity (Joule per degree Kelvin) of a system per unit mass (kilogram). (See also volumetric heat capacity.)

TEMPERATURE: an index of the heat content of materials having the same volumetric heat capacity.

THERMAL CONDUCTIVITY: a measure of the ability of the soil to conduct heat by molecular motion. It is defined as the quantity of heat (Joule) flowing across a unit area (square metre) in unit time (second) in the presence of a temperature gradient of 1°C/m.

THERMAL DIFFUSIVITY: the ratio of the thermal conductivity to the volumetric heat capacity of a substance. It determines the rate at which temperature changes imposed at the surface travel through the profile.

THERMAL RADIATION: see longwave radiation.

TRANSMISSIVITY: the fraction of incident radiation at a given wavelength transmitted through a material.

VOLUMETRIC HEAT CAPACITY: the amount of heat (Joule) required to change the temperature of a unit volume (cubic metre) of soil by 1°C.

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APPENDIX A. Temperature and heat

Temperature and heat are the physical quantities used to describe the thermal regimes of soils. Temperature has units of Kelvins (K) or degrees Celsius (°C), while heat has units of Joules. The heat content of a unit volume (J m^{-3}) of soil depends on: the density of the material (mass per unit volume: kg m^{-3}); the specific heat (the amount of heat [J] required to raise the temperature [K] of a unit mass [kg] by one Kelvin: J kg^{-1} K^{-1}); and the temperature (K) as follows:

$$\text{Heat content} = \text{Density} \times \text{Specific heat} \times \text{Temperature} \qquad (1)$$

The product of the density and the specific heat is known as the volumetric heat capacity. Temperature can therefore be considered as an index of the heat content of materials which have the same volumetric heat capacity. It is sometimes imprecisely referred to as the heat intensity of an object. Temperature is also defined as an expression of the ability with which heat can be transferred to adjacent bodies. This definition is a useful concept because heat is transferred by the process of conduction down temperature gradients from warmer regions to cooler regions.

These concepts can be shown in the following example. Consider the energy contents of equal volumes of air and water, each at a temperature of 20°C (293°K). The specific heats of air and water are 1.01 and 4.18 kJ kg^{-1} K^{-1}, respectively. The densities of air and water are 1.2 and 1000 kg m^{-3}, respectively. The heat contents referenced to absolute zero (0°K) are:

AIR $\quad 1.2 \times 1.01 \times 293 = \quad 351.6$ kJ m^{-3}

WATER $\quad 1000 \times 4.18 \times 293 = \quad 1224740.0$ kJ m^{-3}

Consequently, water contains approximately 3500 times as much thermal energy as the same volume of air at the same temperature. This is one of the reasons why the soil water content has such a large effect on the soil thermal regime. The density, specific heat, and volumetric heat capacity of various soil constituents are given in Table A.1.

TABLE A.1. Density, specific heat, and volumetric heat capacity of various soil materials

Material	Material density (kg m^{-3})	Specific heat (J kg^{-1} K^{-1})	Volumetric heat capacity (MJ m^{-3} K^{-1})[a]
Air	1.204	1010	0.0012
Water	998.2	4182	4.17
Clay (dry)	1600	890	1.42
Clay (40% water)	2000	1550	3.10
Sand (dry)	1600	800	1.28
Sand (40% water)	2000	1480	2.96
Peat (dry)	300	1920	0.58
Peat (40% water)	700	3215	2.25

[a] MJ = MegaJoules = 10^6 Joules.

APPENDIX B. Soil thermal properties

The thermal properties of soil depend on the soil composition, bulk density, and water content (Campbell 1977, 1985; Hanks and Ashcroft 1980). Soil can be considered to be a three-phase material comprising solids (mineral or organic material) and pore space containing air or water. Soil thermal properties vary considerably with changes in water content because the volumetric heat capacity and thermal conductivity of water and air are different. Thermal characteristics of the constituents of soil are presented in Table B.1.

TABLE B.1. Characteristics of soil thermal properties

Property	Symbol	Measures	Consequences
Thermal conductivity	k	Ability of soil to transmit heat. Quantity of heat (J) flowing through a cross-sectional area (m^2) in unit time (s) in the presence of a 1°C temperature gradient	Soils with high k will transmit heat more rapidly under the same temperature gradient
Volumetric heat capacity	C	Amount of heat required to raise the temperature of a unit volume (m^3) of soil by 1°C	Soils with high C will require more heat to cause the same temperature change
Thermal diffusivity	κ	Index of the rate of temperature change with time and soil depth $\kappa = k / C$	Soils with high κ allow surface temperature changes to penetrate rapidly and deeply into the profile
Thermal admittance	A	Index to the variation in the surface soil temperature for a given surface soil heat flux density $A = (k C)^{1/2}$	Soils with high A show less variation in soil temperature amplitude for a given surface heat input
Damping depth	D	Index of the depth to which temperature changes in a given period of time $D = (k \text{ period} / C \pi)^{1/2}$	As D increases, the temperature penetrates to a greater depth

Volumetric Heat Capacity

Soil temperature changes as heat is supplied or withdrawn. The extent of the temperature change depends on both the change in heat storage and the volumetric heat capacity (J m^{-3} K^{-1}) of the soil. The volumetric heat capacity is defined as the amount of heat (J) required to raise the temperature (K) of a unit volume (m^{-3}) by one Kelvin (K). It can be calculated from soil properties as follows:

$$C = x_m C_m + x_o C_o + x_w C_w + x_a C_a \qquad (1)$$

where the subscripts m, o, w and a refer to mineral soil material, organic soil material, water, and air, respectively; C is the volumetric heat capacity of each component (Table B.2); and x is the volume fraction of each component. For practical purposes, the air components can be ignored because the volumetric heat capacity of air is so much less than that of soil or water. The relationship between the volumetric heat capacity, volumetric soil water content, and soil bulk density for a mineral soil is shown in Figure B.1.

TABLE B.2. Thermal properties of natural materials

Material	Density kg m^{-3} (x10^3)	Specific heat J kg^{-1} K^{-1} (x10^3)	Volumetric heat capacity J m^{-3} K^{-1} (x10^6)	Thermal conductivity W m^{-1} K^{-1} (x1)	Thermal diffusivity m^2 s^{-1} (x10^{-6})	Thermal admittance J m^{-2} K^{-1} s$^{-1/2}$ (x10^3)
Still air	0.0012	1.01	0.0012	0.025	20.5	0.005
Turbulent air	0.0012	1.01	0.0012	~125	~107	0.39
Still water (4°C)	1.00	4.18	4.18	0.57	0.14	1.54
Pure ice	0.92	2.10	1.93	2.24	1.16	2.08
Fresh snow	0.10	2.09	0.21	0.08	0.10	0.13
Old snow	0.48	2.09	0.84	0.42	0.40	0.59
Dry sand	1.60	0.08	1.28	0.30	0.24	0.62
Wet sand	2.00	1.48	2.96	2.20	0.74	2.55
Dry clay	1.60	0.89	1.42	0.25	0.25	0.60
Wet clay	2.00	1.55	3.10	1.58	0.51	2.21
Dry peat	0.30	1.92	0.58	0.06	0.10	0.35
Wet peat	1.10	3.65	4.02	0.50	0.12	1.42

Thermal Conductivity

The soil thermal conductivity (W m^{-1} K-1) is a measure of its ability to transmit heat under a given temperature gradient. It is defined as the quantity of heat (Joules) flowing through a unit cross-sectional area (m^{-2}) of soil in unit time (s), a temperature gradient of 1 K/m perpendicular to the surface.

The thermal conductivity can be accurately measured with a number of steady and non-steady state methods. One of these techniques involves making calculations from measured soil thermal gradients and heat flux densities.

A number of different physically based methods have been proposed for estimating the thermal conductivity as a function of soil composition, bulk density, and water content. It turns out to be much more difficult to estimate the thermal conductivity than the volumetric heat capacity this way because the packing and orientation of the soil particles influence the overall conductivity of the soil matrix.

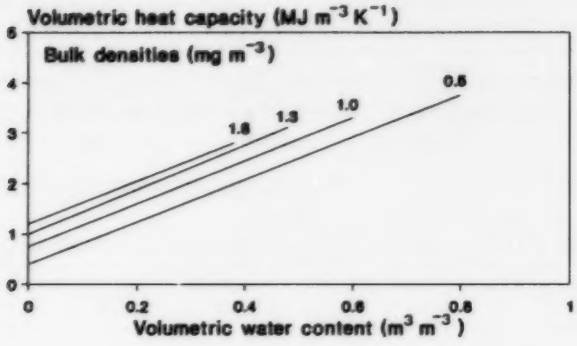

FIGURE B.1. The relationship between the volumetric heat capacity, volumetric water content, and bulk density for a pure mineral soil.

The regression equations proposed by McInnes (1981) can be used to calculate the thermal conductivity of soil as a function of its composition, bulk density, and water content. Thermal conductivities of sand, clay and organic soils as functions of soil water content and bulk density, and calculated using this method, are shown in Figure B.2.

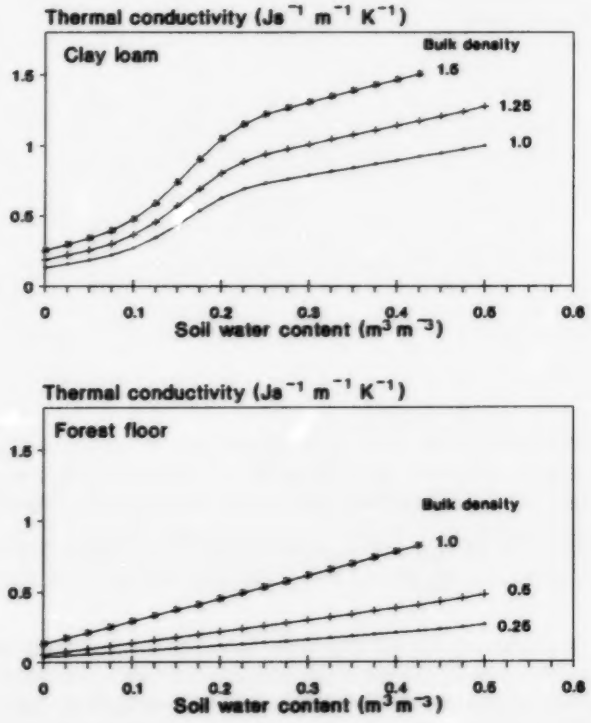

FIGURE B.2. Thermal conductivities of a clay loam, and forest floor organic matter as functions of bulk density and volumetric water content.

Thermal Diffusivity

The thermal diffusivity of soil (m^2 s^{-1}) is the ratio of the thermal conductivity to the volumetric heat capacity. It can be considered to be an index of the time required for temperature changes to travel within the soil profile. Soils with high diffusivities allow changes in surface temperature to penetrate rapidly into the profile. For the same heat input, their temperature regimes are less extreme or variable at the surface than are soils having lower diffusivities. Soils with low diffusivities (dry mineral soil or organic material) concentrate their thermal exchanges with the atmosphere near the ground surface and thus experience relatively extreme diurnal near-surface temperature fluctuations.

Thermal Admittance

The thermal admittance of the soil provides an index of the degree to which the surface temperature changes when heat is added or removed. As the thermal admittance of the soil surface decreases, the variation in surface temperature amplitude increases. It is sometimes referred to as the thermal contact coefficient and is calculated as follows:

$$A = \sqrt{k C} = C \sqrt{\kappa} \qquad (2)$$

where k is the thermal conductivity; C is the volumetric heat capacity; and κ is the thermal diffusivity. Typical admittances for various natural materials are given in Table B.2.

Damping Depth

The damping depth is an index of the depth of penetration of surface temperature changes. It depends on the thermal conductivity, the volumetric heat capacity, and the period (P) over which heat diffuses into the profile. It is calculated as follows:

$$D = \sqrt{2 k / C \omega} \qquad (3)$$

where ω is the angular frequency of the temperature wave ($= 2\pi / P$), $\pi = 3.1416$, and P is 86400 s if the period is one day and k and C have time units of seconds.

The diurnal damping depth for most soils varies from 0.05 to 0.15 m. At 1, 2, 3 and 4 damping depths the diurnal temperature amplitude is reduced to about 37, 14, 5 and 2%, respectively, of the surface temperature amplitude. Consequently, diurnal changes in surface temperature are damped out within the top 0.5 m of the soil profile and annual temperature waves are damped out within about 10 m of the surface. Diurnal and annual damping depths for various soil components are given in Table B.3.

TABLE B.3. Diurnal and annual damping depths of soil, air, and water

Material		Diurnal (m)	Annual (m)
Sand	(dry)	0.08	1.53
Sand	(saturated - porosity 40%)	0.14	2.73
Clay	(dry)	0.07	1.33
Clay	(saturated - porosity 40%)	0.12	2.26
Peat	(dry)	0.05	1.02
Peat	(saturated - porosity 80%)	0.06	1.12
Air	(still)	0.76	14.46
Water	(still - 4°C)	0.06	1.17

APPENDIX C. Soil heat transfer

Heat is simultaneously transferred through soil by the processes of conduction, convection, mass flow of water, and radiation. Conduction is by far the most significant transfer process, although, under certain conditions an appreciable amount of heat can be transferred by convection or mass flow. Convective heat transfer occurs in drying soils, particularly as water evaporates from warm soil pore surfaces and re-condenses on cooler ones. The amount of heat transferred is equal to the mass of water distilled, times the latent heat of vapourization of water. Radiative heat transfer is usually negligible because inter-pore soil temperature gradients are generally small.

Conductive heat transfer is through the net molecular exchange of kinetic energy from more energetic (hotter) to less energetic (cooler) regions. Heat flows down soil temperature gradients, perpendicular to surfaces at the same temperature. This can be illustrated if we consider the soil profile as a series of overlying slabs of unit volume, each having a small finite thickness (z) with a temperature (T). Heat is conducted between adjacent slabs if they are at different temperatures. The amount of heat transferred per unit area of slab (the heat flux density) can be calculated using Fourier's law (Hanks and Ashcroft 1980; Campbell 1985):

$$G = -k\, dT/dz \qquad (1)$$

where G is the flux density of heat (J s^{-1} m^{-2}); dT/dz is the temperature difference divided by the distance between adjacent slabs (K m^{-1}); and k is the thermal conductivity (J s^{-1} m^{-1} K^{-1}) of the slabs. The thermal conductivity describes the ability of a slab to conduct heat. The negative sign is required because heat flows in the direction of decreasing temperature when the heat flux density is positive in the direction of increasing depth.

Heat is either lost, gained, or stored within each slab. The difference between the amount of heat flowing into and out of the slab determines its internal energy, and hence its temperature. The rate of change of internal energy within the slab is a function of the volumetric heat capacity of the slab (J m^{-3} K^{-1}) (the amount of heat required to change the temperature of a slab of unit volume by one K), and the rate of temperature change with time of the slab, i.e.,

$$dG/dz = -C\, dT/dt \qquad (2)$$

where dG/dz is the heat storage rate in the slab; C is the volumetric heat capacity of the slab; and dT is the change in temperature over the time interval dt.

Since the heat storage rate is equal to the difference between the heat flux density into and out of the slab, equations (1) and (2) can be combined to give:

$$dG/dz = d(-k\, dT/dz)/dz = -C\, dT/dt \qquad (3)$$

For soils having uniform thermal properties with depth, equation (3) can be simplified to:

$$dT/dt = \kappa\, (d^2T/dz^2) \qquad (4)$$

where κ (=k/C) is the thermal diffusivity (m^2 s^{-1}) of the soil. A soil with a high diffusivity allows changes in surface temperature to penetrate rapidly into the profile. For the same heat input at the surface, the temperature regime will be less extreme than for a soil with a low diffusivity.

Equations (3) and (4) are parabolic partial differential equations. Simple analytic solutions can be obtained by the assumption that: the soil is infinitely deep (i.e., temperature changes occur to an infinite depth); the soil thermal properties do not change with depth and time and are not temperature dependent; the average temperature over the period of interest is the same at all depths; and the surface temperature varies sinusoidally with time as follows:

$$T(0,t) = T + A_0 \sin(\omega t) \qquad (5)$$

where T(0,t) is the temperature at z=0 (the ground surface) and time t (s); T is the average soil temperature (constant with depth); A_0 is the amplitude of the surface temperature ([maximum-minimum]/2); and ω is the angular frequency of the surface temperature wave oscillation (= 2π radians / period of oscillation). In the case of the diurnal variation, $\omega = (2\pi/86400\ s) = 7.27 \times 10^{-5}$ s^{-1}; for the

annual variation, $\omega = 2 \times 10^{-7}$ s^{-1} or 1.72×10^{-2} d^{-1}. For the lower boundary condition, a constant temperature at some large enough depth is assumed known. This is the average temperature used in equation (5).

The attenuation of the surface temperature amplitude with increasing soil depth depends on the soil thermal properties and the angular frequency of the temperature fluctuation. At a characteristic depth, known as the damping depth, the temperature amplitude decreases to 1/e (i.e., 1/2.7183 = 0.37, or 37% of the surface temperature amplitude). At 2 and 3 damping depths the temperature amplitude is reduced to $A_o e^{-2}$ and $A_o e^{-3}$ (14 and 5%, respectively) of the surface temperature amplitude. The damping depth is calculated from:

$$D = \sqrt{2\kappa/\omega} \qquad (6)$$

For example, the daily and annual damping depths are, respectively, 0.12 and 0.24 m for a moist sand and 0.05 and 1 m for dry organic peat.

Since 95% of the surface temperature amplitude is attenuated within 3 damping depths, the temperature below the 0.36 m depth in a moist sand remains essentially constant throughout the day and only changes slowly over the period of a week. Similarly, the temperature at a depth of 6-7 m is approximately equal to the average annual temperature. In the dry peat soil, the daily temperature only influences the top 0.15 m strongly, and the annual wave, the top 3 m of the profile.

The soil temperature distribution at any time and depth in the profile can be calculated from:

$$T(z,t) = \bar{T} + A_o e^{-z/D} \sin(\omega t[-z/D]) \qquad (7)$$

At any depth z, the amplitude of the temperature fluctuation is smaller than the surface temperature amplitude by a factor of $e^{-z/D}$. A phase shift (time delay of the maximum temperature attained) of z/D radians also occurs at depth z because of heat conduction and storage within the profile. The decreasing amplitude and increasing phase lag with depth are characteristic soil temperature behaviour.

The assumptions of constant soil thermal properties and perfectly sinusoidal temperature behaviour are unrealistic. Forest soils usually have thermal properties which vary considerably with time and depth as functions of changing water content and density. Soil surface temperature variation is also somewhat non-sinusoidal in behaviour because of effects such as day length, vegetative cover, and cloud cover. However, the concept of the temperature damping depth can be useful for predicting the zone of influence in which soil temperatures change over certain periods. Other solution methods which do not have such restrictive assumptions require numerical simulation on a computer to solve the soil heat flow equation (Campbell 1985; Stathers *et al.* 1985, 1988).

APPENDIX D. Calibration of soil temperature sensors

Temperature sensors are calibrated by being immersed in a water bath and compared with a reference temperature sensor over a wide range of temperatures. The water bath should be insulated to stop rapid temperature fluctuations. A thermos flask can be adapted for this application if a temperature-regulated water bath is not available. The water bath should be well mixed and maintained at a constant temperature for a long enough period to allow the sensors to equilibrate with their surroundings (Unwin 1980).

The ice point of water should be used as one of the reference temperatures. A well-mixed ice bath which contains approximately 50% distilled water is reproducible to within 0.01°C. However, a poorly made ice bath may be in error by up to 1°C. Another convenient reference temperature can be achieved by circulating cold tap water through the bath. After a few minutes the tap water temperature usually remains relatively constant.

The reference temperature thermometer should be of good quality, having a high degree of accuracy over the temperature range of interest. Platinum resistance thermometers and specially calibrated mercury-in-glass thermometers can be used. The stem of a mercury-in-glass thermometer must be completely immersed in the bath to minimize differential stem expansion. Parallax error must be avoided when the temperature is read. Sensors should be calibrated at a minimum of 2 and preferably 4 or more temperatures within the temperature range of interest. Calibration accuracy of 0.2°C is usually adequate for most soil thermometers.

APPENDIX E. Installation of soil temperature sensors

The purpose of measuring soil temperature is to characterize the soil thermal regime in the surrounding area. To do this it is necessary to choose a measurement site which represents this area. This involves the choice of a location where the soil surface energy balance is similar to that in the surrounding area and where the subsoil has similar thermal and hydraulic properties to those in the surrounding soil.

Care has to be taken in the installation of the temperature sensor profile. The soil thermal regime can be significantly affected if the profile is not carefully reconstructed. Soil removed from the measurement pit should be carefully stored and refilled in roughly the same order in which it was removed. It should also be re-packed to the same bulk density. The surface should be reconstructed so that the surface energy balance is not affected. These precautions will minimize differences between measured temperatures and those in the surrounding area which the measurements are supposed to represent. It is necessary to replicate the near-surface measurements where greater diurnal variation occurs.

For longer term studies where repeated measurements are required at the same location, the following installation procedures should be used as guidelines to ensure that measured temperatures represent the thermal regime in the surrounding area.

1. Always approach the measurement site from one direction (preferably the downhill side) to minimize disturbance to the heat and water flow fields within the soil. Avoid walking in this area as much as possible so that the surface is not altered. Consider building a removable boardwalk to minimize damage to the soil surface and vegetation, and flag the site boundaries.
2. Excavate a narrow soil pit to the required depth. One face of the pit should be cut with a smooth vertical face. The surface above this face should remain undisturbed.
3. Soil which is removed from the pit should be stored on a tarpaulin so that the profile can be reconstructed in roughly the same order in which the soil was removed.
4. Label the sensor lead wires appropriately with the depth and other necessary information prior to installation. The labels should be durable and weather resistant.
5. Define the surface, the reference point used to determine lower profile depths. Starting from the bottom of the profile and working toward the surface, repeat the following procedure at each measurement depth:
   a. Drill a 100-150 mm deep hole, the diameter of the sensor, horizontally into the vertical face of the soil profile.
   b. Push the sensor into the hole, ensuring that there is good thermal contact with the surrounding soil.
   c. Bury about 0.2 m of lead wire at the same depth as the sensor to minimize heat conduction toward the sensor.
   d. Refill the soil pit to the depth of the next sensor, taking care to reconstruct the soil profile.
6. Ensure that the centre of the sensor is at the correct depth, particularly in the top 0.2 m of the profile where large temperature gradients occur.
7. Ensure that the disturbed surface is reconstructed to look like the surrounding area.

8. Cover the sensor leads or put them in a protective box to avoid animal damage. Ensure that the lead wires are also located such that the surface energy and water balances are not affected.

Disturbance of the soil profile is not as critical for point measurements in a soil pit. The sensor should be installed into a shaded face of the soil pit to minimize the effects of radiative heating. It should also be placed horizontally to avoid vertical soil temperature gradients. The sensor should be in good thermal contact with the surrounding soil and left in place long enough to equilibrate with the surrounding temperature before the measurement is recorded.

APPENDIX F. Temperature scales and conversions

Temperature scales are derived from the fundamental laws of thermodynamics. They are calibrated with reference to fixed points relating to easily reproducible changes of state of various materials. Three commonly used reference points are the ice, triple, and steam points of water. The ice point is defined as the temperature at which ice melts under an atmospheric pressure of 101.3 kPa (760 mm Hg). The triple point (0.01°C at 6.106 kPa) is the only equilibrium state where ice, air-saturated water, and water-saturated air co-exist. The steam point is defined as the temperature of the vapour of boiling water at a pressure of 101.3 kPa (Omega 1984).

The relationships between temperature scales for these reference points are given in Table F.1. The Kelvin and Rankine scales have their zeros at absolute zero, but have different scale units. The scale units of the Kelvin and Celsius scales are identical but differ from the Rankine and Fahrenheit scales. The scale units of the Rankine and Fahrenheit scales are also identical. In relative terms, 9° Rankine or Fahrenheit is equal to 5° Kelvin or Celsius or 4° Réaumur. Temperatures are usually reported in degrees C for biological application. The following equations and the reference points given in Table F.1. are used for making conversions between scales.

$$°F = 9/5\ °C + 32.0 \quad (1)$$

$$°C = 5/9\ °F - 32.0 \quad (2)$$

$$K = 273.15 + °C \quad (3)$$

TABLE F.1. Temperature scales

Scale	Symbol	Ice point	Triple point	Steam point
Kelvin	K	273.15	273.16	373.15
Celsius	°C	0.00	0.01	100.00
Fahrenheit	°F	32.00	32.02	212.00
Rankine	°R	491.67	491.69	671.67
Réaumur	°R	0.00	0.01	80.00